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CONTENTS TO VOLUME 247

NUMBER 1

USE OF THE UNIVERSAL STAGE IN SEDIMENTARY PETROGRAPHY *Charles M. Gilbert and Francis J. Turner* 1

A NOTE ON TWO ASPECTS OF THE GEOCHEMISTRY OF CARBON *G. Evelyn Hutchinson* 27

A PRELIMINARY STUDY OF THE FLORA OF THE BRANDON LIGNITE .... *E. S. Barghoorn and William Spackman, Jr.* 33

THE GALLIUM CONTENT OF IGNEOUS ROCKS .... *E. B. Sandell* 40

Dendraster elsmerensis, N. SP. .... *J. Wyatt Durham* 49

DISCUSSION

NATIONAL RESEARCH COUNCIL COMMITTEE ON THE INTERRELATIONS OF PLEISTOCENE RESEARCH *Chauncey D. Holmes* 63

REVIEWS

Atomic Energy; Norman Wait Harris Lectures; by KARL K. DARROW 64
Electrical Measurements in Theory and Application; by A. W. SMITH *L. W. McKeehan* 64

Very High Frequency Techniques, Vols. I and II; by THE STAFF OF THE RADIO RESEARCH LABORATORY, HARVARD UNIVERSITY *William L. Briscoe* 65

Heat Conduction, with Engineering and Geological Applications; by L. R. INGERSOLL, ET AL. .... *W. W. Watson* 66

General Chemistry; by FRANCIS E. RAY; 66
Organic Chemistry; by FRANCIS E. RAY .... *Elliott Pierce* 67

Frontiers in Chemistry. Vol. V, by R. E. BURK and OLIVER GRUMMITT *Henry C. Thomas* 67

A Textbook of Geomorphology; by PHILIP G. WORCESTER *Sidney E. White* 68

A Hadrosaurian Dinosaur from New Jersey; by E. H. COLBERT *Richard S. Lull* 68

Farm Soils, Their Management and Fertilization; by E. L. WORTHEN; 69
Soil Physics; by L. D. BAVER .... *C. L. W. Swanson* 70

Human Geography, An Ecological Study of Society; by C. L. WHITE and G. T. RENNER .... *Stephen B. Jones* 70

PUBLICATIONS RECENTLY RECEIVED 71

NUMBER 2

THE FLOOR OF THE SHENANDOAH VALLEY . . . *Philip B. King* 73

LATE PLEISTOCENE AND RECENT CHANGES OF SEA LEVEL ALONG THE COAST OF SANTA BARBARA COUNTY, CALIFORNIA . . . *J. E. Upson* 94

THE DISTRIBUTION OF ORGANIC MATTER IN SEDIMENTS FROM THE NORTHERN RED SEA . . . *A. F. Mohamed* 116

Mooreocrinus Bowsheri, NEW SPECIES FROM THE CHESTER SERIES OF NORTHEASTERN OKLAHOMA *Harrell L. Strimple* 128

Hybocrinus Crinerensis, NEW SPECIES FROM THE ORDOVICIAN OF OKLAHOMA *Harrell L. Strimple and William T. Watkins* 131

TYPES AND HYPODIGMS . . . *Norman D. Newell* 134

REVIEWS

Outlines of Physical Chemistry; by FARRINGTON DANIELS . . . *P. A. Lyons* 143

Précis des Découvertes Somiologiques ou Zoologiques et Botaniques; by C. S. RAFINESQUE . . . *Charles L. Remington* 144

College Physics; by HENRY A. PERKINS . . . *C. G. Montgomery* 144

NUMBER 3

ISOSTASY AND THE ORIGIN OF SIAL AND SIMA AND OF PARENTAL ROCK MAGMAS .... *Walter Wahl* 145

PROCESSES OF EROSION ON STEEP SLOPES OF OAHU, HAWAII *Sidney E. White* 168

THE POLAROGRAPHIC METHOD FOR DETERMINING TRACE ELEMENTS IN ROCKS AND MINERALS .. *Esther W. Claffy* 187

DISCUSSION AND COMMUNICATIONS

INTERNATIONAL CONFERENCE ON QUATERNARY RESEARCH ... 200

CHAMBERS'S MINERALOGICAL DICTIONARY ... *Edward S. Smith* 200

REVIEWS

Geological Explorations in the Island of Celebes; Geological Summary and Petrology; under the leadership of H. A. BROUWER *John Rodgers* 201

Illustrated Catalogue of North American Devonian Fossils: Unit 1-E, part a, Family Auloporidae; edited by ERWIN C. STUMM *Carl O. Dunbar* 205

A Monograph of the Existing Crinoids; Vol. 1, by A. H. CLARK *Carl O. Dunbar* 205

The Corridor of Life; by W. E. SWINTON .... *Joseph T. Gregory* 206

Vibration and Sound; by PHILIP M. MORSE .... *Richard F. Humphreys* 207

PUBLICATIONS RECENTLY RECEIVED 208

NUMBER 4

METASOMATIC GRANITIZATION OF BATHOLITHIC DIMENSIONS. PART 1 .... *Peter Misch* 209

REGULARITY OF OLD FAITHFUL GEYSER, YELLOWSTONE NATIONAL PARK, WYOMING .... *Philip F. Fix* 246

RADIOACTIVITY OF OCEAN SEDIMENTS. VI. CONCENTRATIONS OF THE RADIOELEMENTS IN MARINE SEDIMENTS OF THE SOUTHERN HEMISPHERE ... . *William D. Urry* 257

Dolichoharpes AND THE ORIGIN OF THE HARPID FRINGE *Harry B. Whittington* 276

REVIEWS

The Essentials of Organic Chemistry; by C. D. PORTER and T. D. STEWART *Alfred H. Frye* 286

Manual of Coastal Delineation from Aerial Photographs; by P. G. McCURDY *Sidney E. White* 286

PUBLICATIONS RECENTLY RECEIVED 288

NUMBER 5

STUDIES OF SUPERGLACIAL DEBRIS ON VALLEY GLACIERS
Robert P. Sharp 289

THERMODYNAMIC RELATIONS OF IMMISCIBILITY AND CRYSTALLIZATION OF MOLTEN SILICATES
Tom. F. W. Barth and Terkel Rosenqvist 316

A NEW SPECIES OF *Paleorhinus* FROM THE TRIASSIC OF TEXAS
Wann Langston, Jr. 324

REVIEWS

Cybernetics; by NORBERT WIENER .... *Henry Margenau* 342

Practical Spectroscopy; by GEORGE R. HARRISON, RICHARD C. LORD and JOHN R. LOOFBOUROW .... *W. W. Watson* 344

The Interpretation of Spectra; by WM. M. VENABLE .. *Richard F. Humphreys* 345

Organic Reactions. Vol. IV. Edited by ROGER ADAMS ET AL. ..*Alfred H. Frye* 346

The Chemical Constitution of Natural Fats; by T. P. HILDITCH ....*S. L. Liu* 347

When the Earth Quakes; by JAMES B. MACELWANE ......*Chester R. Longwell* 348

Fundamentals of Physical Science; by KONRAD B. KRAUSKOPF .. *John Rodgers* 349

Crystal Structures; by RALPH W. G. WYCKOFF ....*Horace Winchell* 350

A Source Book in Greek Science; by MORRIS R. COHEN and I. E. DRABKIN
Henry Margenau 351

PUBLICATIONS RECENTLY RECEIVED 352

NUMBER 6

THE LACCOLITHIC SERIES ...... *O. T. Jones and W. J. Pugh* 353

METASOMATIC GRANITIZATION OF BATHOLITHIC DIMENSIONS PART II. STATIC GRANITIZATION IN SHEKU AREA, NORTHWEST YUNNAN (CHINA) .... *Peter Misch* 372

SPHEROIDAL WEATHERING OF IGNEOUS ROCKS *Randolph W. Chapman and Mildred A. Greenfield* 407

REVIEWS

Introduction to Historical Geology; by RAYMOND C. MOORE *Lawrence Whitcomb* 430

PUBLICATIONS RECENTLY RECEIVED 431

NUMBER 7

FABRIC CHANGES IN YULE MARBLE AFTER DEFORMATION IN COMPRESSION, PART 1 . .. ... *Eleanora Bliss Knopf* 433

TWO ENDOCRANIAL CASTS OF CETACEANS FROM THE OLIGOCENE OF NEW ZEALAND . . . . *B. J. Marples* 462

STRATIGRAPHY AND TRILOBITE FAUNAL ZONES OF THE GARDEN CITY FORMATION, NORTHEASTERN UTAH
Reuben J. Ross, Jr. 472

GROWTH PATTERNS IN THE CERATOPSIA
Richard Swann Lull and Stephen Wood Gray 492

OBITUARY

MAX BLANCKENHORN . . . .*Rushdi Said* 504

REVIEWS

The Chemistry and Technology of Enzymes; by HENRY TAUBER
Harold G. Cassidy 506

Dynamic Aspects of Biochemistry, American Edition; by ERNEST BALDWIN
Harold G. Cassidy 506

Annual Report on the Progress of Chemistry for 1947 (Vol. 44)
Harold G. Cassidy 507

Microwaves and Radar Electronics; by ERNEST C. POLLARD and JULIAN M. STURTEVANT . . . .*Sanborn C. Brown* 507

Submarine Geology; by FRANCIS P. SHEPARD . . . .*John B. Lucke* 509

Foraminifera: Their Classification and Economic Use; by JOSEPH CUSHMAN
Karl M. Waage 511

Bibliography of the Literature on Sodium and Iodine in Relation to Plant and Animal Nutrition . . . .*Eva M. Low* 511

PUBLICATIONS RECENTLY RECEIVED 512

NUMBER 8

THE BASE OF THE CAMBRIAN IN THE SOUTHERN APPALACHIANS, PART I .... *Philip B. King* 513

THE FAULT AT THE WEST EDGE OF THE TRIASSIC IN SOUTHERN PENNSYLVANIA .... *George W. Stose* 531

FABRIC CHANGES IN YULE MARBLE AFTER DEFORMATION IN COMPRESSION, PART II .... *Eleanora Bliss Knopf* 537

COMPRESSION TESTS ON ORIENTED SPECIMENS OF YULE MARBLE *Henry A. Lepper, Jr.* 570

QUANTITATIVE AREAL GEOLOGY OF THE UNITED STATES *Donald V. Higgs* 575

REVIEWS

The Atmospheres of the Earth and Planets; by GERARD P. KUIPER *Rupert Wildt* 584

The Face of the Moon; by RALPH B. BALDWIN .... *Rupert Wildt* 585

The Sphere of Sacrobosco and Its Commentators; by LYNN THORNDIKE *Dirk Brouwer* 586

Abriss Der Geologie; by PROF. DR. ROLAND BRINKMANN ...... *Ernst Cloos* 586

Introduction to Atomic Physics; by OTTO OLDENBERG .... *H. Margenau* 588

Foundations of Modern Physics; by THOMAS B. BROWN ...... *H. Margenau* 588

The Science and Engineering of Nuclear Power. Vol. II; by CLARK GOODMAN .... *W. W. Watson* 588

Botanik der Gegenwart und Vorzeit in Culturhistorischer Entwickelung; by KARL F. W. JESSEN .... *Robert Bloch* 589

The Avian Egg; by ALEXIS L. ROMANOFF and ANASTASIA J. ROMANOFF *D. Rudnick* 591

PUBLICATIONS RECENTLY RECEIVED 592

NUMBER 9

PREFERRED ORIENTATION OF CALCITE IN YULE MARBLE *Francis J. Turner* 593

THE BASE OF THE CAMBRIAN IN THE SOUTHERN APPALACHIANS. PART II. . . . .*Philip B. King* 622

ANDEAN UPLIFT AND EROSION SURFACES NEAR UNCIA, BOLIVIA. . . . . *Eugene H. Walker* 646

THE JOURNAL CHANGES EDITORS. 664

REVIEWS

The Structure of Matter; by FRANCIS OWEN RICE and EDWARD TELLER *Henry C. Thacher, Jr.* 666

Sedimentary Rocks; by F. J. PETTIJOHN..................*John Rodgers* 667

Map of the Americas ....*John Rodgers* 668

Klockmann's Lehrbuch der Mineralogie; by PAUL RAMDOHR *Horace Winchell* 668

Minerals and How to Study Them; by the late EDWARD S. DANA *Walden P. Pratt* 669

Animals Without Backbones: An Introduction to the Invertebrates; by RALPH BUCHSBAUM ....*Karl M. Waage* 670

Recent Marine Shell-Bearing Mollusca of the South Sea Islands (Part 1); by KOTORA HATAI ....*P. A. Morris* 671

Tellus, a quarterly journal of geophysics 672

NUMBER 10

METASOMATIC GRANITIZATION OF BATHOLITHIC DIMENSIONS, PART III. RELATIONSHIPS OF SYNKINEMATIC AND STATIC GRANITIZATION . . . . *Peter Misch* 673

ORIGIN OF PIMPLE MOUNDS . . . . *E. L. Krinitzsky* 706

CERAMIC PRODUCTS; THEIR GEOLOGICAL INTEREST AND ANALOGIES . . . . *W. O. Williamson* 715

REVIEWS

Fear, War and the Atom Bomb; by P. M. S. BLACKETT *Henry Margenau* 750

Quantum Mechanics; by L. I. SCHIFF . . . . *Henry Margenau* 751

Principles of Petroleum Geology; by CECIL G. LALICKER . . *John E. Sanders* 751

PUBLICATIONS RECENTLY RECEIVED 752

NUMBER 11

GRANITE AND METASOMATISM .... *Reginald A. Daly* 753

UNUSUAL STYLOLITES ... *Daniel B. Blake and Chalmer J. Roy* 779

SPHEROIDAL STRUCTURES IN ARIZONA VOLCANICS
John H. Feth and John W. Anthony 791

THE ULTRABASIC ROCKS OF MIDDLE RIVER RANGE, B. C.
H. W. Little 802

DEFINITIONS OF ARKOSE .... *Steven S. Oriel* 824

REVIEWS

Two Problems of Marine Geology: Atolls and Canyons; by PH. H. KUENEN
Dwight R. Crandell 830

Gesteine und Minerallagerstätten. Erster Band: Allgemeine Lehre von den Gesteinen und Minerallagerstätten; by PAUL NIGGLI ... *Charles B. Sclar* 831

ERRATA 832

NUMBER 12

A PRE-WISCONSIN FOREST SUCCESSION IN THE PUGET LOWLAND, WASHINGTON
Henry P. Hansen and J. Hoover Mackin 833

RADIUM CONTENT OF ULTRAMAFIC IGNEOUS ROCKS: II. GEOLOGICAL AND CHEMICAL IMPLICATIONS
G. L. Davis and H. H. Hess 856

PLEISTOCENE STRATIGRAPHIC SEQUENCE IN NORTHEASTERN KANSAS .... *John C. Frye and A. Byron Leonard* 883

TWO NEW SPECIES OF *Acrocrinus* FROM THE PENNSYLVANIAN OF OKLAHOMA .... *Harrell L. Strimple* 900

REVIEWS

PUBLICATIONS RECENTLY RECEIVED 905

INDEX TO VOLUME 247 906
CONTENTS, VOLUME 247

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USE OF THE UNIVERSAL STAGE IN SEDIMENTARY PETROGRAPHY

CHARLES M. GILBERT AND FRANCIS J. TURNER

ABSTRACT. The universal stage is of particular value in the investigation of certain sedimentary petrographic problems. Customary universal stage technique is readily applied to sedimentary petrography, but the following special applications are recommended and discussed: goniometric procedure for the determination of habit and interfacial angles in idiomorphic and sub-idiomorphic minerals, investigation of mutual relations of cementing minerals or secondary growths, determination of feldspar composition and recognition of authigenic feldspar, distinction between quartz and clear plagioclase, and identification of calcite and dolomite by optical methods.

INTRODUCTION

THE purpose of this paper is to draw the attention of sedimentary petrographers to some of the uses of the universal stage in routine petrographic examination of sedimentary rocks. As a result of their own experience, the writers believe that a universal stage is a piece of apparatus essential for a modern sedimentary laboratory and offer the following suggestions for its use.

1) One good petrographic microscope should be reserved for this work. On it a universal stage should be mounted permanently to avoid the waste of time otherwise involved in repeatedly setting up centering and adjusting the instrument prior to each investigation.

2) For all routine measurements of such properties as optic orientation, optic sign, axial angle, extinction angles, and interfacial angles, whether in uniaxial or biaxial minerals, a stage with three rotation axes (innermost vertical, north-south horizontal, and east-west horizontal) is thoroughly satisfactory. Our preference is for the three-axis stage of Silge and Kuhne, San Francisco, or the standard four-axis stage of Leitz.

3) The classic Federow procedure described in such standard German works as those of Berek (1924), Reinhard (1931) and Nikitin (1936) is recommended. Various aspects and modifications of the Federow method will be found in a number of papers written in English (Barber, 1936, pp. 221-289; Nemoto, 1938; Haff, 1941; Turner, 1940, 1942, 1947), and

in addition, a clear, concise outline of the essential procedure has been given by Haff (1942). The Federow method involves plotting all measured directions and planes on a stereographic or an equal-area projection net. In the writers' opinion this is an advantage rather than otherwise, because a clear picture of the optic and crystallographic orientation of the measured mineral develops as measurement proceeds, and any error in measurement or identification of a given direction is readily perceived and may be corrected. Moreover, use of a projection net is essential in identifying crystal faces and zones, and is the simplest method of measuring interfacial angles with a universal stage. Those unfamiliar with the use of a projection net in this connection are referred to a paper by Haff (1940) which gives a clear account of the essential procedures involved.

4) Material adapted to investigation with a universal stage includes thin sections and grain preparations permanently mounted in plastic, Canada balsam, hyrax, or some other resin.

5) The beginner is reminded that the chief advantage of the universal stage is that certain properties (e.g., axial angle, optic orientation, intercleavage angles) may be measured directly in a single crystal; whereas with an ordinary petrographic microscope such properties are usually determined by comparing optical reactions observed in a number of crystals presumed to be identical in composition. But the precision of the universal stage is still limited by the degree of accuracy with which a position of extinction or the sharpness of a cleavage surface can be estimated visually. Most measurements involve errors of the order of one or two degrees; so that the universal stage, although a convenient instrument, yields data in many cases little more precise than those determined by ordinary petrographic means.

GENERAL TECHNIQUE APPLIED TO SEDIMENTARY PETROGRAPHY

In sedimentary, just as in igneous or metamorphic petrography, the most generally useful optical data for the rapid measurement of which the universal stage is specially suited are, 1) the positions of the indicatrix axes X, Y, and Z in relation to visible crystallographic planes and directions and 2) the optic axial angle and optic sign. It is not within the scope of this paper to describe the Federow procedure for obtaining these data, but the use of this procedure is invaluable in the solution of problems that commonly arise in sedimentary petrographic work.

One of the special difficulties of sedimentary petrography arises from the great variety of relatively rare minerals that are likely to be present in heavy concentrates. Such, for example, are lawsonite, diaspore, scheelite, spodumene, etc. Where only one or two grains are available for study in a single mount, optic axial angle and optic sign are properties whose determination may be essential for identification of the minerals. Also with respect to some of the more abundant minerals found in sedimentary rocks optical data obtained by means of a universal stage may be invaluable, as, for example, in the precise determination of feldspar composition (Turner, 1947), the distinction between orthorhombic and monoclinic pyroxene by the relation of Z to the prismatic cleavage, the identification of varieties of monoclinic pyroxene by means of optic axial angle, and the distinction between apatite and colorless andalusite lying on the front pinacoid {100} cleavage.

Strongly colored minerals or minerals with strong dispersion are not suited to investigation by means of a universal stage because in such minerals it is impossible to locate precisely the positions of the indicatrix axes. This applies particularly to mounted grains (as contrasted with thin sections) since the effects of dispersion and absorption are increased by the relatively great thickness of grain. Epidote minerals, dumortierite, pumpellyite, and many amphiboles belong to this category.

Universal stage procedure is also essential in those special investigations in which measurement of preferred orientation of common constituents of sedimentary rocks (notably quartz, carbonates, and micas) is desirable (Ingerson, 1944, p. 640).

DETERMINATION OF CRYSTAL HABIT AND INTERFACIAL ANGLES

So many sedimentary minerals, especially those of heavy concentrates, occur as idiomorphic or subidiomorphic crystals, that crystal habit is an important diagnostic character. Although the prevailing habit of a given mineral may usually be judged by inspection of a number of typical grains with the ordinary microscope, it is advantageous to be able to tilt individual grains about an axis within the plane of the microscope stage. This is readily done if the slide is mounted on a universal stage. The amount of additional detailed information readily available through this simple procedure alone is significant and lends perspective to other observations made with the ordinary microscope. Such a procedure may be particularly desirable in cases where variation in habit or in mean dimensions of grains is investigated statistically in problems relating to paleogeography (Smithson, 1939, pp. 423-427; Hagerman, 1933).

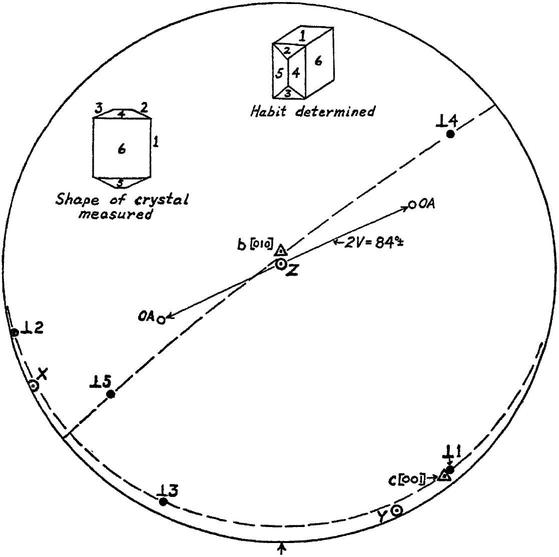

Fig. 1. Equal-area projection of heulandite (Example 1).

Approximate determination of interfacial angles and of zonal relations of faces bounding idiomorphic crystals is another technique specially useful in sedimentary petrography, both as a step towards identifying minerals and in interpretation of rock fabrics as observed in thin sections. The position of any well developed crystal face (or cleavage plane) is readily located within one or two degrees by bringing it into the vertical north-south or east-west plane. The interfacial angles are measured between the poles of (normals to) faces plotted on the projection net and are subject to errors of several degrees, but the data obtained are in many cases sufficiently accurate to allow certain identification of both the mineral and the measured faces. Marked differences between refractive index of the mineral and that of the mounting medium (e.g., zircon in Canada balsam) increases the error owing to total reflection at surfaces adjoining the measured face.

Example 1.—The identification of minute crystals of authigenic heulandite in an upper Miocene sandstone (Santa Margarita formation) from Santa Cruz County, California (U.C. #51-8), is cited as an illustration of the use of interfacial angles and habit in identifying sedimentary minerals. The mineral was suspected of being a zeolite because of its low birefringence (.005±) and low refractive index (mean R. I. = 1.485 ± .003), but positive identification was not possible by ordinary microscopic means.

Figure 1 shows the indicatrix and poles of measured crystal faces of a single crystal plotted on the lower hemisphere of an equal-area projection net. The measurements made with a universal stage from which these points were plotted are[1]:

X	25 → 1
Y	115 → 2 ↓ 45 ↑ 39
⊥ face #1	310 ← 5
⊥ face #2	13 → 1
⊥ face #3	63 → 5
⊥ face #4	40 ← 21
⊥ face #5	36 → 24

---

[1] For all readings given in this paper the following convention is used. The first figure is the reading on the scale for the inner vertical axis of the stage; the arrow and second figure indicate the direction and amount of downward tilt (on the N.-S. axis). To plot these on the lower hemisphere of the projection net the following rules may be followed. Turn the zero position of the transparent overlay to the measured angle (first figure) indicated on the circumference of the net; plot the point at the angle of tilt (second figure) measured in from the circumference of the net along the east-west axis in the direction indicated by the arrow. In the case of measurement of Y, the third and fourth figures in the reading are the direction and corresponding amount of tilt around the horizontal east-west axis necessary to bring an optic axis into the vertical position. The optic axes are plotted on the net, along the great circle 90 degrees from Y, in the direction and amount indicated, starting at the horizontal east-west axis of the net.

The position of Z is located graphically at 90 degrees from both X and Y.

Examination of the plotted points reveals that the poles of three faces (#1, #2, and #3) all lie on a great circle (the dashed line through these three poles), and also that this great circle passes approximately through X and Y. Thus within the limits of accuracy to be expected in such measurements, all three of these faces are co-zonal and are normal to the plane including X and Y. The corresponding zone axis (labeled b{010} in Figure 1) and the optically determined point Z should thus coincide. Also these two points are very close to the great circle which includes the poles of faces #4 and #5. The two great circles are thus mutually perpendicular within the limits of experimental error. These data suggest strongly that the mineral is either monoclinic or orthorhombic.

The general habit of the crystal can be clearly seen by tilting it into various positions, and the interfacial angles can be read from the projection net. Comparison with published data (Dana, 1899, pp. 570-610) indicates that two zeolites, mordenite and heulandite, have the habit of the mineral in question; but the interfacial angles identify it as heulandite, with the following forms developed: #1 = {001}, #2 = {201}, #3 = $\{\bar{2}01\}$, #4 and #5 = $\{1\overset{+}{\bar{1}}0\}$, and #6 = {010}. The following table gives the interfacial angles measured and, for comparison, the angles reported by Dana (shown in italics to the nearest degree).

	∧ ⊥ 2		∧ ⊥ 3		∧ ⊥ 4		∧ ⊥ 5	
⊥ 1	63	*64*	66	*66*	88	*89*	89	*89*
⊥ 2			50	*50*	34	*33*	32	*33*
⊥ 3					35	*33*	32	*33*
⊥ 4							45	*44*

Interfacial angles and the identification of faces on the single crystal illustrated in this example have been checked, with an accuracy of two or three degrees, by measurements made on many other crystals from the same concentrate. The precision of these measurements is dependent on the sharpness with which the positions of crystal faces may be estimated visually, so that the operator can decide at the time of measurement which readings are likely to be most reliable. Obviously some crystals are more favorable for such measurements than others, and in general crystals with the most clearly defined faces should be chosen. The crystal chosen for this example is particularly favorable, but because it lay on its side pinacoid {010}, face #6, other grains must be measured to identify that face positively. Such measurements prove that face #6 is indeed the side pinacoid {010} and that it is strictly normal to Z.

In minerals of very low birefringence, such as the zeolites, relatively slight changes in refractive index (possibly caused by slight variations in composition or changes in temperature) may alter the optic sign and orientation or produce large changes in the optic axial angle. Crystal habit and interfacial angles are probably much more constant than such optical data, and are therefore valuable in identification of such minerals. In the present case, for example, the optical character of the heulandite immersed in Canada balsam and heated enough to harden the balsam is very different from the optical character observed when the mineral is mounted in plastic (Castolite) hardened without heating. (The optical orientation reported in this example is for the unheated mineral.) The heating necessary to harden the Canada balsam caused the optic sign to be reversed and the positions of X and Z to be interchanged. The interfacial angles and habit of the mineral are identical in the two media.

MUTUAL RELATIONS OF CEMENTING MINERALS AND SECONDARY OUTGROWTHS

The universal stage is particularly adapted to establishing the existence of secondary outgrowths on detrital mineral grains and to bringing out the mutual relations of cement minerals and authigenic minerals in general. Contrary to what might be expected, a random section across a grain partly bounded by sharply defined crystal faces in many cases yields irregular outlines, so that the existence of crystal faces is not suspected until these have been tilted into a vertical plane. For example, in mounted light fractions of some samples of the Santa Margarita sandstone of southern Santa Cruz County, California, most of the numerous grains of feldspar appear to be angular and irregular. Yet tilting in the appropriate direction on a universal stage reveals the surprising fact that approximately half of these grains are at least partly bounded by crystal faces, and in many grains these can be shown to be clear secondary outgrowths developed around somewhat cloudy detrital grains originally irregular.

During the examination of thin sections of some Eocene reservoir sands from the Kettleman Hills-Coalinga area, California, the writers found that, in general, quartz and feldspar grains appear to be irregularly bounded where they project into or adjoin cavities. Here, too, appropriate tilting of these free edges, even through ten or fifteen degrees, reveals the presence of sharp secondarily developed crystal faces that would otherwise remain undetected. By measuring and plotting several such faces and one or more optical directions in any grain, it is commonly possibly to identify the individual faces and so to obtain a picture of the exact crystal form that tends to develop by secondary enlargement of the detrital grains. Moreover, it may be possible to obtain reliable information as to the order of development of outgrowths and cementing minerals during lithification of the sediments in question (Gilbert, 1948). Several instances are cited below as examples of this technique.

Example 2.—This illustration of the identification of crystal forms developed on a secondary outgrowth of quartz is taken from the Eocene McAdams sand of the Kettleman Hills area, California (U.C. #S3/3C). Figure 2 shows the optic axis and poles of faces measured on a secondarily enlarged quartz grain. In the inset sketch representing the general form of the grain measured (Q_1), stippled areas are detrital quartz and diagonal lines indicate secondary outgrowths on several detrital grains; cross-hatched areas are detrital feldspar. The sketch is highly idealized to indicate the number and relative positions of the faces on several outgrowths, but it should be understood that at any one tilt of the stage only one or two of these faces are clear. With no tilt at all, faces #1 and #4 alone are distinct.

With the universal stage the positions of the optic axis (c axis) and the pole of each suspected crystal face are measured in the usual manner and with the following results:

Optic Axis	39 ← 20	⊥ face #4	350 → 3
⊥ face #1	348 → 4	⊥ face #5	28 → 32
⊥ face #2	260 ← 20	⊥ face #6	258 ← 19
⊥ face #3	315 → 16	⊥ face #7	317 → 14

Each of these readings is plotted on the lower hemisphere of an equal-area (or stereographic) projection net, and the angle between the optic axis and the pole of each face is read from

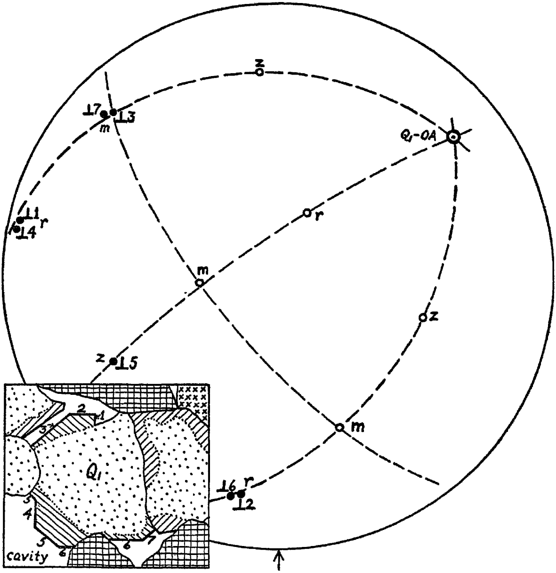

Fig. 2. Equal-area projection of quartz (Example 2).

the net. These angles are then compared with published angles for quartz to give a tentative identification of the various faces, as follows:

	∧ Optic axis	Suggested form	Correct angle (to nearest degree)
⊥ face #1	56°	r $\{10\bar{1}1\}$ or z $\{01\bar{1}1\}$	52°
⊥ face #2	56°	r $\{10\bar{1}1\}$ or z $\{01\bar{1}1\}$	52°
⊥ face #3	90°	m $\{10\bar{1}0\}$	90°
⊥ face #4	53°	r $\{10\bar{1}1\}$ or z $\{01\bar{1}1\}$	52°
⊥ face #5	54°	r $\{10\bar{1}1\}$ or z $\{01\bar{1}1\}$	52°
⊥ face #6	54°	r $\{10\bar{1}1\}$ or z $\{01\bar{1}1\}$	52°
⊥ face #7	92°	m $\{10\bar{1}0\}$	90°

These results and an inspection of the points plotted on the projection suggest that among the seven faces three forms are represented; namely #3, #7 = $\{10\bar{1}0\}$; #1, #4 and #2, #6 = $\{10\bar{1}1\}$; #5 = $\{01\bar{1}1\}$. Alternatively #1, #4 and #2, #6 could be identified with $\{01\bar{1}1\}$ and #5 with $\{10\bar{1}1\}$. The above identification can be checked by two independent means, 1) the zone relationships of faces and 2) interfacial angles.

Zone relations: The poles of all prism faces must lie on a great circle (zone) 90 degrees from the optic axis. This great circle is drawn on Figure 2 and includes face #3 while face #7 is only slightly off the zone. Assuming that face #3 is the common unit prism $\{10\bar{1}0\}$, the poles of the other unit prism faces can be located on the same great circle by measuring 60 degrees on either side of face #3 and are shown as open circles labeled m. Now if the other faces are indeed unit rhombohedrons, each must be co-zonal with one of the prism faces and the optic axis. This condition is satisfied in this case as shown by the diagram where each of the faces in question lies on or very near one of these great circles. For reference, the poles of the other three faces of positive and negative unit rhombohedron forms are shown by open circles labeled r and z.

Interfacial angles: The angle between each pair of faces is measured from the projection net and compared with the correct interfacial angles for the forms identified. The results are as follows, where the correct angle is given in italics:

	∧ 2 (r)	∧ 3 (m)	∧ 4 (r)	∧ 5 (z)	∧ 6 (r)	∧ 7 (m)
1 (r)	89 *86*	34 *38*	2 *0*	47 *46*	89 *86*	32 *38*
2 (r)		65 *67*	89 *86*	48 *46*	2 *0*	66 *67*
3 (m)			36 *38*	67 *67*	67 *67*	2 *0*
4 (r)				47 *46*	87 *86*	34 *38*
5 (z)					46 *46*	66 *67*
6 (r)						68 *67*

Within the limits of accuracy obtainable with the universal stage, the angles check and the faces can be considered as satisfactorily identified.

Because the angles between optic axis and each of the various faces are independent of the interfacial angles, and both are independent of the zone relations, identifications checked by all three methods can be considered certain. Where only two or three faces occur on a crystal the identification is possible in the same manner although it may be less convincing than with a larger number; but if a single face is present the only character available is the angle between the optic axis and the face, in which case identification is considered tentative or inconclusive.

It should also be apparent that the identification of crystal faces in this manner refers to crystal forms only and not to individual faces. Thus the positive (r) and negative (z) unit rhombohedrons in quartz cannot be distinguished. The writers follow the convention that if only one unit rhombohedron form is present it is called the positive one (r); if both are present either may be called positive (r) and the other negative (z).

Example 3.—The relationship between calcite and secondary quartz outgrowths in the Eocene McAdams sand of the Kettleman Hills area (U.C. #S3/7C) is illustrated by Figure 3. The inset sketch is highly idealized and illustrates a portion of a thin section showing a triangle of calcite (C) surrounded by a detrital quartz (stippled) with secondary outgrowths (diagonal lines). The cross-hatched areas are feldspar. The optic axes of two quartz grains (Q_1 and Q_2) and of calcite (C), together with the poles of faces #1, #2, #3 and the pole of a calcite cleavage, were located in the usual manner with the universal stage and plotted on the lower hemisphere of the equal-area projection net. The readings made with the stage were as follows:

Calcite optic axis	200 ← 56	⊥ face #1	196 ← 26
Quartz$_1$ optic axis	254 ← 24	⊥ face #2	126 ← 34
Quartz$_2$ optic axis	195 ← 65	⊥ face #3	339 → 13
Calcite cleavage	182 ← 14		

The angle between the optic axis of calcite and the pole of the calcite cleavage as measured from the projection net is 43 degrees. Since this angle agrees as well as can be expected with the correct angle (44½°) between the c axis and rhombohedral cleavage of calcite, it can be assumed that the fracture plane measured is indeed the rhombohedral cleavage $\{10\bar{1}1\}$. With the pole of this cleavage and the optic axis (c axis) plotted on the projection, the positions of the other common rhombohedron faces of calcite can be located and are shown by open triangles labeled r ($\{10\bar{1}1\}$) and e ($\{01\bar{1}2\}$). None of these approximates the positions occupied by the poles of faces #1 and #2, so that clearly neither of these faces is a common rhombohedron of calcite.

The angles between the optic axis of quartz$_1$ and the poles of faces #1 and #3 are 52 degrees and 89 degrees respectively,

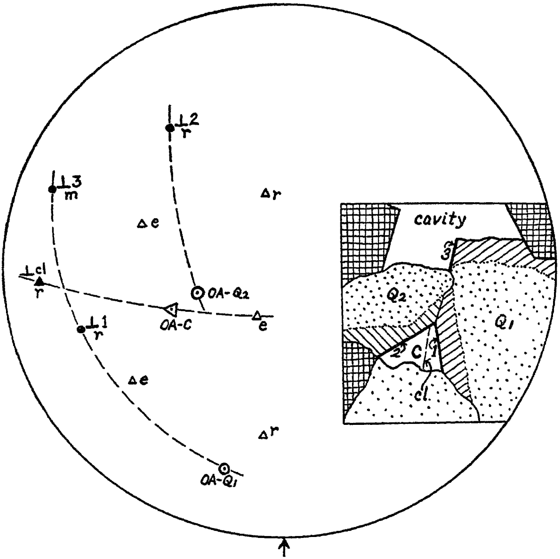

Fig. 3. Equal-area projection of calcite, C, and quartz, Q_1 and Q_2 (Example 3).

which agree with the correct angles for the unit rhombohedron r ($\{10\bar{1}1\}$) and unit prism, m ($\{10\bar{1}0\}$) of quartz. Also these two faces are co-zonal with the optic axis (c axis). Thus face #1 is determined as the rhombohedron of quartz$_1$ and is not a rational face of calcite. The optic axis of quartz$_2$ is inclined to the pole of face #2 at an angle of 51 degrees, a value that agrees with identification of face #2 as the unit rhombohedron, r $\{10\bar{1}1\}$, of quartz$_2$. This determination appears reasonable but is not positive since only one face is to be found on quartz$_2$ and zonal relations and interfacial angles cannot be used as a check on the identification.

It is concluded that faces #1 and #2 are unit rhombohedrons on quartz outgrowths, Q_1 and Q_2 respectively, and are not rational faces of calcite as at first they would appear to be. Therefore, the calcite must have crystallized later than the quartz outgrowths and has filled a void bounded by previously developed faces of secondary quartz. This conclusion is borne out by similar observations made throughout this and other thin sections from the same formation. The same technique can be used to investigate the boundary between two different quartz outgrowths, such as the one shown in the inset sketch of Figure 3. Such boundaries generally approximate to plane surfaces; but results obtained by the writers indicate that most of them are not rational faces of either grain of quartz. This was the result in the case illustrated here and probably indicates that the two quartz outgrowths developed simultaneously.

Reference data.—For convenient reference the following crystallographic angles of quartz and calcite are given to the nearest degree.

TABLE I

Common crystallographic angles of quartz

c	axis	Λ	⊥m	$(10\bar{1}0)$	90°
c	axis	Λ	⊥r	$(10\bar{1}1)$	52°
c	axis	Λ	⊥z	$(01\bar{1}1)$	52°
⊥m	$(10\bar{1}0)$	Λ	⊥r	$(10\bar{1}1)$	38°
⊥m	$(01\bar{1}0)$	Λ	⊥z	$(01\bar{1}1)$	38°
⊥m	$(01\bar{1}0)$	Λ	⊥r	$(10\bar{1}1)$	67°
⊥m	$(10\bar{1}0)$	Λ	⊥z	$(01\bar{1}1)$	67°
⊥r	$(10\bar{1}1)$	Λ	⊥r	$(\bar{1}101)$	86°
⊥z	$(01\bar{1}1)$	Λ	⊥z	$(1\bar{1}01)$	86°
⊥r	$(10\bar{1}1)$	Λ	⊥z	$(01\bar{1}1)$	46°

TABLE II

Angles to rhombohedron faces of calcite

c	axis	Λ	⊥r	$(10\bar{1}1)$	45°
c	axis	Λ	⊥e	$(01\bar{1}2)$	26°
⊥r	$(10\bar{1}1)$	Λ	⊥r	$(\bar{1}101)$	75°
⊥e	$(01\bar{1}2)$	Λ	⊥e	$(\bar{1}012)$	45°
⊥r	$(10\bar{1}1)$	Λ	⊥e	$(\bar{1}012)$	71°
⊥r	$(10\bar{1}1)$	Λ	⊥e	$(01\bar{1}2)$	38°

DETERMINATION OF COMPOSITION OF FELDSPAR

For methods of determining the composition of twinned grains of feldspar as seen in thin sections the reader is referred to Nikitin (1936) and Turner (1947).

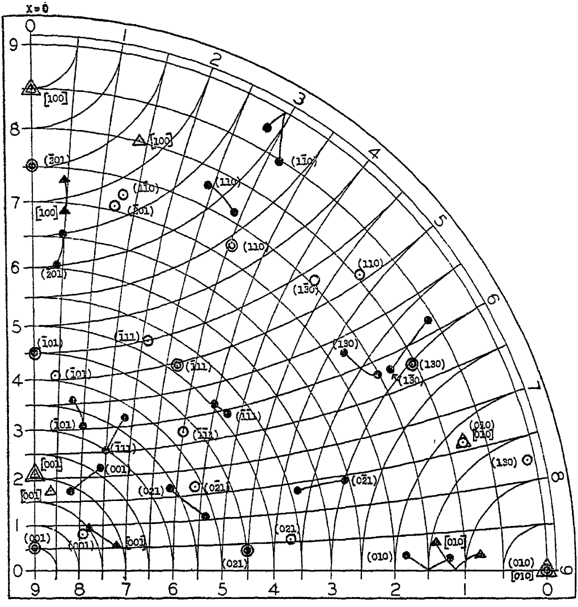

Fig. 4. Inclination of X, Y, and Z to axes and poles of faces in orthoclase, microcline, and albite. Triangles are axes; circles are poles of faces. Solid black points are albite (Ab_{100}-Ab_{90}) with label at pure albite; single circles and triangles are microcline; double circles and triangles are orthoclase.

A special case, of great importance in problems of diagenesis and cementation, is the determination of narrow rims (outgrowths) of secondary feldspar surrounding detrital feldspar grains, and of minute idiomorphic crystals of authigenic feldspar. In such grains or outgrowths, whether seen in thin section or in individual grain mounts, it is commonly possible to locate approximately with a universal stage the positions of a number of crystal faces and the directions X, Y, and Z. These are plotted on a projection net (as in previous examples) and the three angles between the pole of each face and X, Y, and Z respectively are measured from the net and recorded. These angles are plotted from the appropriate corners of the stereographic diagram (after Nikitin, 1936, Pl. VII)reproduced as Figure 4, each set of three angles defining a point on the diagram. Even with the most favorable conditions and reliable measurements, the points thus obtained

TABLE III

Interfacial angles in feldspar to the nearest degree

		Orthoclase Microcline	Albite
$(001) \wedge (010)$	zone of a axis	90	86°24′
$(001) \wedge (\bar{1}01)$	zone of b axis	50	52
$(001) \wedge (\bar{2}01)$		80	82
$(010) \wedge (1\overset{+}{\bar{1}}0)$	zone of c axis	59+	60+
$(010) \wedge (1\overset{+}{\bar{3}}0)$		29+	30+
$(001) \wedge (1\overset{+}{\bar{1}}0)$	co-zonal	68	$\wedge (110) = 65$; $\wedge (1\bar{1}0) = 69$
$(001) \wedge (\bar{1}11)$		55	$\wedge (\bar{1}11) = 56$; $\wedge (\bar{1}\bar{1}1) = 58$
$(\bar{1}11) \wedge (1\bar{1}0)$		57	55
$(\bar{1}01) \wedge (\bar{1}11)$	co-zonal	27	26
$(010) \wedge (\bar{1}\overset{+}{\bar{1}}1)$		63	$\wedge (\bar{1}11) = 60\frac{1}{2}$; $\wedge (\bar{1}\bar{1}1) = 66$
$(\bar{1}11) \wedge (\bar{1}\bar{1}1)$		54	53
$(\bar{2}01) \wedge (110)$	co-zonal	46	46
$(\bar{2}01) \wedge (\bar{1}11)$		39	39
$(110) \wedge (\bar{1}11)$		85	85
$(001) \wedge (1\overset{+}{\bar{3}}0)$		78	$\wedge (130) = 74$; $\wedge (1\bar{3}0) = 80$
$(\bar{1}01) \wedge (1\overset{+}{\bar{1}}0)$		69	$\wedge (110) = 70$; $\wedge (1\bar{1}0) = 66$
$(\bar{1}01) \wedge (1\overset{+}{\bar{3}}0)$		78	$\wedge (130) = 80$; $\wedge (1\bar{3}0) = 74$
$(\bar{2}01) \wedge (1\overset{+}{\bar{3}}0)$		67	$\wedge (130) = 68$; $\wedge (1\bar{3}0) = 62$
$(\bar{1}11) \wedge (130)$		78	$\wedge (130) = 77$; $\wedge (1\bar{3}0) = 52$
$(\bar{1}\bar{1}1) \wedge (130)$		54	$\wedge (130) = 57$; $\wedge (1\bar{3}0) = 82$
$(110) \wedge (\bar{1}\bar{1}1)$		57	57
$(1\bar{1}0) \wedge (1\bar{1}1)$		84	81

are not likely to coincide with labeled points representing the principal faces of feldspar, but by their proximity to labeled points they will serve to identify the faces measured and the nature of the feldspar. This identification can then be checked by means of interfacial angles and zone relations between faces. Particularly useful in this regard is the common face (and cleavage plane) {010}, since the angles between its pole and X, Y, and Z respectively vary significantly with the composition of the feldspars. Likewise, if the positions of any of the three crystallographic axes (a[100], b[010], c[001]) can be located on the projection net, the three angles between each of these axes and X, Y, and Z respectively can be plotted on Figure 4 and compared with the labeled axial points as a check on the identification of the feldspar in question. The axes are particularly useful in this connection because their mutual relations to X, Y, and Z vary considerably in the different feldspars.

Reference data.—Crystallographic and optical data used in determining low temperature feldspars by this method are

TABLE IV

Angles between X, Y, and Z and poles of common feldspar faces to nearest degree (After Nikitin, 1936, pp. 94-101)

	Orthoclase			Microcline			Albite (Ab_{100})			Albite (Ab_{90})		
	X	Y	Z	X	Y	Z	X	Y	Z	X	Y	Z
⊥ (010)	90	90	0	74	82	18	88	72	18	88½	78	12
⊥ (001)	85	5	90	82	13	79½	68	26	76	73	19	82
⊥ (110)	36	72	59½	49½	80½	42	28	78	66	33	76	61
⊥ $(1\bar{1}0)$	36	72	59½	22	72½	77	32	85	58½	28½	87½	61½
⊥ $(\bar{1}01)$	45	45	90	49	41	86	60	32	80	54½	36	82
⊥ (021)	86	45	45	85	53	37	73½	34	61	79½	38	54
⊥ $(0\bar{2}1)$	86	45	45	73½	38	56½	75½	64½	30	76	56	37
⊥ $(\bar{2}01)$	15	75	90	23	70½	78	30	60½	86	25	65	85
⊥ (130)	63	80	29½	78	89	12	57½	72½	38	62	75	32½
⊥ $(1\bar{3}0)$	63	80	29½	46½	75½	47	62	77	31½	58½	84	32
⊥ $(\bar{1}11)$	50½	51	63	45	52	69	65	30	75	58½	37½	72
⊥ $(\bar{1}\bar{1}1)$	50½	51	63	62½	43	60	61	51	52½	58½	50½	55
c [001]	69	21	90	73	17½	87	85	19	72	81	15	78
a [100]	5	85	90	17½	80	76	22	69	85	17½	73	85
b [010]	90	90	0	74	82	18	86½	76	14	88	82	8

tabulated in Tables III and IV. The interfacial angles cited are measured between the normals to (poles of) faces in question. Within the limits of error involved in universal-stage measurements, the angle between any two faces may in most cases be regarded as identical for feldspar of any variety or composition.

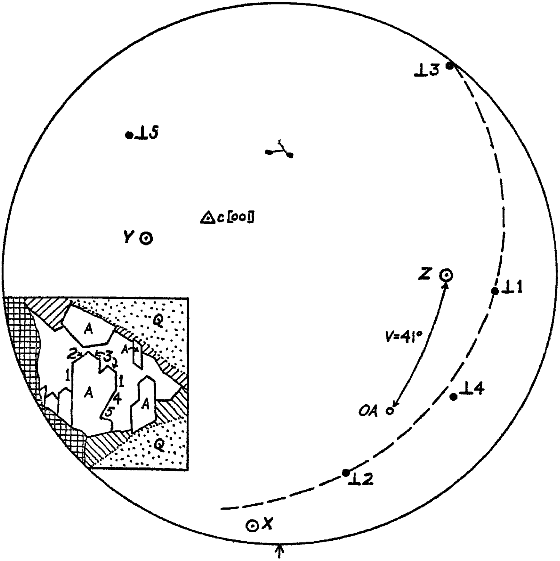

Fig. 5. Equal-area projection of albite (Example 4).

Example 4.—Authigenic albite from the Eocene McAdams sand of the Kettleman Hills area (U.C. #S3/10C) is illustrated by the equal area projection and inset sketch reproduced as Figure 5. In the sketch, the stippled areas are detrital quartz (Q) with secondary outgrowths shown by diagonal lines; cross-hatched areas are detrital feldspar, and euhedral crystals (A) are albite projecting into a cavity. The shapes of the albite crystals and the faces developed on them and secondary quartz are not apparent unless the section is tilted; nor are they all apparent at any one position of tilt.

Measurements of X, Y, and Z, an optic axis, and the poles of all visible faces, made with a universal stage on the largest albite crystal, are as follows:

X	84 → 6
Y	345 → 48 ↓ 29
Z	0 ← 40

⊥ face #1	356 ← 24
⊥ face #2	288 ← 24
⊥ face #3	52 ← 2
⊥ face #4	324 ← 24
⊥ face #5	317 → 26

Plotted on the lower hemisphere of a equal-area projection net, these readings appear as the points shown in Figure 5. The three angles between X, Y, and Z and the pole of each face are read from the projection net and the faces tentatively identified by plotting these angles on the diagram reproduced as Figure 4, with the following results:

	∧X	∧Y	∧Z	Crystal form represented
⊥ face #1	88	72	17	(010) of Ab_{100}
⊥ face #2	28	88	61	$(1\bar{1}0)$ of Ab_{91}
⊥ face #3	34	76	60	(110) of Ab_{90}
⊥ face #4	60	75	34	(130) of Ab_{95} or $(1\bar{3}0)$ of Ab_{100}
⊥ face #5	61	31	77	$(\bar{1}01)$ or $(\bar{1}11)$ of Ab_{100}

If the first four faces are correctly identified they must lie on a single great circle since they are all in the zone of the c axis (see Table III); as can be seen in Figure 5 this condition is satisfied. A further independent check on the identity of the faces can be made by determining the interfacial angles from the projection net and comparing these with the proper angles given in Table III. Face #5 in particular is not certainly identified by the angles between its pole and X, Y, and Z respectively, which indicate that it may be either $((\bar{1}01)$ or $(\bar{1}11)$; but the ambiguity is removed by referring to the interfacial angles. These prove that face #5 is the unit pyramid $(\bar{1}11)$. Similarly angles to faces #2 and #3 prove that face #4 is $(1\bar{3}0)$, not (130). The following tabulation shows the measured interfacial angles and, for comparison, the correct angles in italics.

	Δ⊥#2 $(1\bar{1}0)$		Δ⊥#3 (110)		Δ⊥#4 $(1\bar{3}0)$		Δ⊥#5 $(\bar{1}11)$		$(\bar{1}01)$
⊥ #1 (010)	62	*60*	59	*60*	29	*30*	62	*61*	*86*
⊥ #2 $(1\bar{1}0)$			60	*59*	33	*30*	58	*55*	*66*
⊥ #3 (110)					87	*89*	85	*85*	*70*
⊥ #4 $(1\bar{3}0)$							50	*57*	*80*

With the faces thus certainly identified, the c axis [001], can be located on the projection net since it is parallel to faces #1, #2, #3, and #4; that is, it is the pole of the great circle through the poles of these faces. Located as it is from four faces considered conjointly, the position of the c axis [001], is probably more reliable than the position of any single face. Its inclination to X, Y, and Z plotted on Figure 4 agrees closely with that of pure albite.

It is concluded that the idiomorphic feldspar crystal in question is nearly pure, untwinned albite having a Baveno habit. The smaller crystals present in the same cavity are similar but they have fewer faces developed. The accuracy obtained in these measurements is about all that can be expected with the universal stage, but great precision is not required for satisfactory identification, particularly if the angular relation between X, Y, and Z and one of the pinacoids or one of the crystallographic axis is known.

By the use of these methods and similar ones applied to quartz (see examples 2 and 3), the relationship between this albite and the secondary quartz outgrowths in contact with it can be determined. Quartz is molded around idiomorphic albite (particularly the upper left part of the sketch in Figure 5), but the albite crystals appear to have grown on top of thin quartz outgrowths. These latter bottom contacts of albite against secondary quartz are not rational faces of either mineral, and it is concluded that the two crystallized more or less simultaneously or that their periods of crystallization overlapped.

Example 5.—Authigenic feldspars are fairly numerous among the light minerals in the Santa Margarita sandstone, southern Santa Cruz County, California. Figure 6 is an equal-area projection of a crystal of authigenic orthoclase from this formation (U.C. #S1/29).

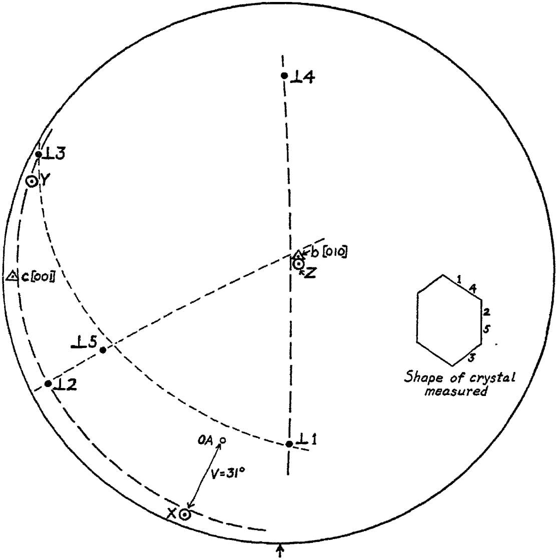

Fig. 6. Equal-area projection of orthoclase (Example 5).

The following measurements made with the universal stage were plotted on the projection net (Figure 6):

X	69 → 4
Y	339 → 4 ↓ 55
⊥ face #1	93 → 36
⊥ face #2	26 → 7
⊥ face #3	333 → 2
⊥ face #4	89 ← 28
⊥ face #5	24 → 30

The position of Z is plotted 90 degrees from X and Y. The angles between X, Y, and Z and the pole of each of the five faces is read from the projection net, and the probable form of each face is determined by plotting these angles on the diagram reproduced as Figure 4, with the following results:

	∧X	∧Y	∧Z	Crystal form represented
⊥ face #1	39	74	56	{110} of orthoclase
⊥ face #2	43	47	89	{1̄01} of orthoclase
⊥ face #3	85	6	88	{001} of orthoclase
⊥ face #4	37	71	60	{110} of orthoclase
⊥ face #5	49	49	66	{1̄11} of orthoclase

An independent check on this identification of the faces may be obtained by measuring the interfacial angles on the projection net and comparing them with the correct interfacial angles for orthoclase (Table III). The agreement is good as shown by the following tabulations where the correct angles are italicized.

	∧⊥#2 (1̄01)		∧⊥#3 (001)		∧⊥#4 (110)		∧⊥#5 (1̄11)	
⊥ #1 (11̄0)	68	*69*	68	*68*	61	*61*	57	*57*
⊥ #2 (1̄01)			53	*50*	70	*69*	23	*27*
⊥ #3 (001)					66	*68*	56	*55*
⊥ #4 (110)							85	*84*

Another independent check is obtained by using the b and c crystallographic axes. Since faces #1 and #4 are two faces of the unit prism, the c axis [001] is the normal to the great circle passing through the poles of these two faces; likewise the b axis [010] is the normal to the great circle passing through the poles of faces #2 and #3, (1̄01) and (001) respectively. The angles between each of these axes and X Y, and Z are read from the projection net and plotted on Figure 4. Furthermore, the angles between each crystallographic axis and the poles of faces (other than those used to locate the axis) can be measured and compared with the correct angles for an additional check if necessary. In the present case, the agreement is good between all of these angles and the corresponding correct angles for orthoclase.

The zone relations between poles of faces, crystallographic axes, and indicatrix axes should also be checked in connection with this identification, particularly where combinations of three or more co-zonal faces are present. In the present example, faces #1 (11̄0), #3 (001), and #5 (1̄11) should be co-zonal if they have been correctly identified; Figure 6 shows that they are. Likewise, faces #2 (1̄01) and #5 (1̄11) and the b axis [010] should be and are co-zonal within the limits of error of measurement. Also if faces #2 and #3 respectively are ($\bar{1}01$) and (001) of orthoclase, the great circle containing their poles should include X, Y, and the c axis [001], a condition which is satisfied in Figure 6.

RAPID DISTINCTION BETWEEN QUARTZ AND CLEAR FELDSPAR

Statistical investigation of the relative amounts of the principal light constituents in sandstones involves rapid distinction between alkali feldspars (with refractive indices below that of Canada balsam), plagioclase (oligoclase or a more calcic variety), and quartz. Clear grains of plagioclase without obvious twinning are not uncommon and are difficult or impossible to distinguish quickly from quartz by ordinary means. The universal stage is an efficient device for this purpose. Its use has previously been recommended by Doeglas (1940), but the procedure outlined by him, namely, that required for general distinction of uniaxial from biaxial minerals, is unnecessarily complex. The problem involved is the particular case of distinguishing a *positive* uniaxial mineral (quartz) from a biaxial mineral with large optic axial angle (feldspar), and can be solved simply and quickly as follows:

1) Set the north-south and east-west rotation axes of the stage near the zero positions. With the grain in question illuminated between crossed nicols, insert the gypsum plate and note whether the reaction is one of compensation or addition. Remove the plate and rotate on the innermost vertical axis of the stage so as to bring the grain into extinction with its slow direction (Z′) parallel to the north-south axis.

2) Tilt through any angle on the east-west axis of the stage. If the grain becomes illuminated (passes out of extinction), it is not quartz. If it stays in extinction for all positions of tilt on the east-west axis, it is probably quartz; but the same effect would result from a grain of feldspar oriented with X or Y parallel to the east-west axis of the stage—a most unlikely possibility.

3) To eliminate this latter possibility, tilt to any new angle on the north-south axis, and rotate on the innermost vertical axis until the grain is again in extinction with its slow direction (Z′) parallel to the north-south axis. If the grain again remains dark for all positions of tilt on east-west axis, it is uniaxial and positive (i.e., it is quartz).

If the grain is oriented with the optic axis nearly perpendicular to the plane of the slide, the test may be more decisive if the gypsum plate is retained in position throughout. Extinction of the grain is then indicated not by darkness, but by the unmodified first order red color of the plate.

In practice the writers have found that a procedure involving steps #1 and #2 alone will yield reliable results in statistical work. The possible misidentification of quartz mentioned in step #2 is unlikely, but in any case the error arising from this cause is probably less than the sampling error in grain counts of 100 to 200 grains. Also the procedure described is used in practice only if a grain cannot be identified quickly by other means (refractive index, cleavage, or twinning). Carried out in this way the operation is very nearly as rapid as any other microscopic method involving counting, and is considerably more reliable.

OPTICAL DISTINCTION BETWEEN CALCITE AND DOLOMITE

Emmons (1943, p. 189) has described an optical method of distinguishing calcite from dolomite, based upon the difference in the refractive index of the extraordinary ray (epsilon) in these two minerals (1.50 in dolomite; 1.486 in calcite). With the north-south and east-west axes of the stage at zero, a section of carbonate bounded by Canada balsam is brought to extinction with the fast direction (e′) parallel to the north-south axis. The analyser is removed, and the section is tilted on the east-west axis until the refractive index of the carbonate exactly matches that of Canada balsam. Both the angular position around the inner vertical axis and the angle of tilt on east-west are read in order to locate e′ (for which R.I. = 1.540) on a projection net. The position of the optic axis (e) is now located in the usual way and plotted on the projection. The angle e Λ é (optic axis Λ é) is 29½ degrees in the case of dolomite, or 36 degrees for calcite. Smaller angles than 29½ degrees indicate ankeritic dolomite (e.g., 24½ degrees for dolomite in which there is 20% replacement of Mg by Fe).

The writers have found that a modification of this method may sometimes be applied in thin sections to grains of carbonate adjoining quartz—a condition commonly encountered where a sandstone is cemented by calcite or dolomite. The procedure is as follows:

1) Locate the respective optic axes in the carbonate grain (e_c) and in the adjoining quartz grain (e_q). On the projection net draw the great circles respectively normal to the two optic

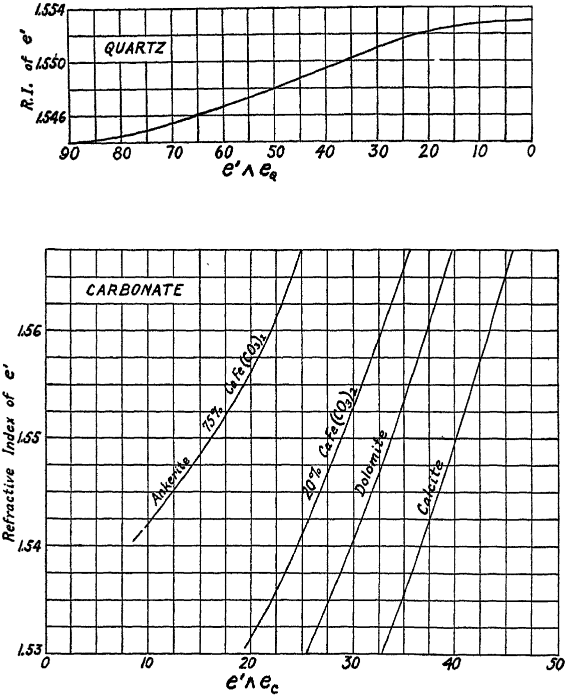

Fig. 7. Refractive index of é in quartz and carbonate.

axes. These great circles intersect in a point representing a common line lying in the basal planes of both crystals, and which is therefore a possible ordinary ray direction for both quartz and carbonate.

2) With the east-west axis at zero, rotate the section on the innermost vertical axis and tilt on the north-south axis so as to bring the common ordinary ray direction parallel to the east-west axis. (The appropriate angles of rotation and tilt are first read from the projection). If these two operations have been performed accurately, both grains will remain in extinction for any angle of tilt on the east-west axis.

3) With the analyser removed, tilt on the east-west axis until the refractive indices of the quartz and carbonate exactly match. Light in both crystals is now vibrating north-south in a common direction (e'), which is plotted on the projection net. The angle between e' and the optic axis of quartz (e_q) is now read from the projection net and the refractive index of e' determined from the upper curve of Figure 7. This index of refraction plotted against the angle between e' and the optic axis of carbonate (e_c) on the lower set of curves in Figure 7 should identify the carbonate.

Quartz-carbonate pairs in which one of the two grains is oriented with its optic axis nearly normal to the plane of the section are most favorable for measurement, since only a small angle of tilt is then necessary to bring the common ordinary ray direction parallel to the east-west axis (step 2). The degree of reliability of this method, whether as originally described by Emmons or in the modified form just presented, depends mainly on the accuracy with which the refractive index of a mineral may be matched with that of the adjacent medium. In the writers' experience the range of determined values in the angle of $e_c \Lambda e'$ in grains of a single carbonate mineral is in some cases too great for reliable identification of the mineral as calcite or dolomite. In other cases where clear-cut boundary surfaces between the media were readily observed and a good research microscope was used, the results were found to be consistent and distinction between calcite and dolomite was made with some confidence. Distinction of ankeritic dolomite from calcite is more reliable on account of the higher refractive index of the former mineral as compared with pure dolomite.

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A NOTE ON TWO ASPECTS OF THE GEOCHEMISTRY OF CARBON

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ABSTRACT. Examination of the best data for the C^{12}/C^{13} ratio indicates that the carbon of the sediments as a whole, when any reasonable ratio of shales to limestones is assumed, contains less C^{13} than does the carbon of meteorites. The rather meagre comparable data give means for meteoritic and for igneous carbon that differ significantly. If Rankama is correct in assuming that the meteoritic ratio characterizes the earth as a whole, there must have been a slight non-biological impoverishment of C^{13} in the superficial part of the lithosphere. It is, however, equally likely that the meteorites do not provide a reliable terrestrial standard. Rankama's figures for the ratio in the carbon of *Corycium* are significantly different from those for igneous carbons, but not from those for meteoritic carbons. His conclusion that *Corycium* is an Archaean fossil is only valid if some difference between meteoritic carbon and that of the accessible lithosphere be assumed.

The available data for methane production by domestic animals and for the populations of such animals indicates at least 45 million metric tons of methane are liberated into the atmosphere per year. There are many other terrestrial sources of methane and a figure of 90 million metric tons may well be too low. This represents about 0.06% of the carbon circulating in the photosynthetic cycle. Mechanisms of oxidation of CH_4 must operate, either photochemically in the stratosphere or biologically in the soil.

THE two aspects of the geochemistry of carbon discussed below, though not closely related, are conveniently presented together.

I. THE TERRESTRIAL $C^{12}/^{13}$ RATIO

In two papers of great importance Rankama (1948a, b) has examined all the available data, including a number of new determinations, on the ratio of C^{12} to C^{13} in meteoritic and terrestrial materials. The marked difference between the carbon of sedimentary carbonates and the reduced carbon of bituminous sediments and recent and fossil plants, first noticed by Nier and Gulbransen (1939) is firmly established, and is used by Rankama as an argument that the problematic structures known as *Corycium enigmaticum* Sederholm from the Bothian phyllites of Finland are actually genuine Archaean fossil plants. This argument, though subject to slight modification, is almost certainly correct. There is, however, one aspect of the treatment of the data that demands further consideration.

Using only the data obtained by Murphey (Murphey and

Nier, 1941, Rankama 1947a) and by Rankama himself, which data are known to be instrumentally uniform and so comparable, the following ranges and mean values are available,

Meteorites (7 determinations)	89.8-92.0	mean *91.26*
Igneous graphite and calcite (4 determinations)	89.4-90.2	mean *89.82*
CaCO$_3$ sediments (13 determinations)	88.4-89.5	mean *89.15*
Carbonaceous sediments (18 determinations) including *Corycium*)	90.3-92.9	mean *91.88*
Known recent and fossil plants (29 determinations	91.3-93.1	mean *92.02*
Corycium carbon (3 determinations)	90.8-92.0	mean *91.4*

Statistical examination, using the *t*-test (Fisher 1934) for the significance of two means indicates that,

1.) The meteoritic and igneous carbons show a highly significant difference ($P<0.01$)

2.) The calcareous sediments and the igneous carbons show a highly significant difference ($P<0.01$)

3.) The bituminous sediments and the igneous carbons show a highly significant difference ($P<0.01$)

4.) The bituminous sediments and the meteoritic carbons show a difference of very moderate significance ($P = 0.03$)

5.) *Corycium* carbon by itself is not significantly different from meteoritic carbon ($P = 0.8$) but the difference between *Corycium* carbon and igneous carbon does have a high degree of significance ($P<0.01$)

The bituminous and carbonaceous sediments and the plant carbons are obviously concordant and a weighted mean of 91.97 is therefore the best value for the reduced carbon of the biosphere.

Rankama believes that the best estimate of the ratio for the earth as a whole is obtained from meteorites. As Wickman (1941) and independently Kamen (1946) pointed out, the values of the ratios for oxidized and reduced carbon in the sediments must be related to the total means of such oxidized and reduced carbon, the quantity of carbon in the atmosphere and hydrosphere being negligible. If the mean of the available data for the meteorites is taken as a standard, the divergences from this standard are

oxidized C (CaCO$_3$)	reduced C (plant and bitumen)
—2.11	+0.71

These divergences imply about three times as much reduced as oxidized carbon in the sediments, a conclusion which is quite inconsistent with the rest of the available geochemical evidence.

The argument introduced by Wickman may, however, be reversed. Taking as an extreme estimate (Kuenen 1941 for continental sediments), the limestones as comprising 15% of the sediments, containing 11.35% carbonate carbon, and the shales 65% of the sediments, containing 0.81% carbon (cf. Clarke 1924, Kuenen 1941, Conway 1942, Hutchinson 1944), the ratio of oxidized carbon to reduced carbon will be 1.703 : 0.526 or 3.24 : 1. To obtain such a ratio the original material, separated on the one hand into carbonate $C^{12}/C^{13} = 89.15$), on the other into reduced organic carbon ($C^{12}/C^{13} = 91.97$) would have had an isotopic ratio of 89.82, or precisely that obtained from a consideration of the rather meagre material derived from the accessible igneous rocks. Alternatively taking the more usual estimate of 5% limestones and 80% shales (Clarke 1924) which perhaps underestimates the limestones, the same reasoning leads to a figure for the original ratio of 90.66. The true value probably lies between these two extremes. It is evident that if Rankama is correct in his belief that the meteoritic ratio is also the ratio for the earth as a whole, some prebiological impoverishment of C^{13} has occurred in the superficial layers of the lithosphere, in the production of the magmas which, on solidification, produced the igneous carbons, and in the production of the volcanic gases from which the sedimentary carbons are ultimately derived. The fact that neither the igneous carbons nor the volcanic gases need be primary is irrelevant. The mass relations show unequivocally that the meteoritic mean cannot apply to the original material of the accessible lithosphere and of the biosphere. If the meteoritic mean applies to the whole earth, then clearly a non-biological process produced slight impoverishment in C^{13} in the superficial layers at an earlier stage than the separations involved in calcium carbonate precipitation and photosynthetic fixation.

As an alternative to the hypothesis that a non-biological separation occurred in the earth's crust, it may be urged that the meteoritic ratio is not inherently likely to be applicable to the earth. At any given temperature the mean molecular velocity of any gaseous compound of C^{13} will be slightly less than that of the analogous compound of C^{12}. Since it is reasonably certain from the high cosmic and low terrestrial abundance of the element that the earth lost a considerable amount of carbon early in the history of its cooling, the mean terrestrial value of the ratio C^{12}/C^{13} presumably is a function, among other things, of the mass of the earth. An identical ratio therefore would be expected only if the meteorites were derived by fragmentation of a planet of the same mass as the earth. While the elegant researches of Brown and Patterson (1948) have greatly increased the probability that the meteorites are in fact the products of the fragmentation of a terrestrial planet, it would be premature to assume that in such a planet the C^{12}/C^{13} ratio could not have differed from that of terrestrial carbon. The differences observed between igneous and meteoritic carbon may therefore depend on original differences established during the cooling of the earth and of the hypothetical planet. Further determinations of the isotopic ratio in carbon from meteorites and from igneous rocks are obviously urgently needed. In particular a study of the distribution of ratios in a long series of meteorites might produce information of considerable importance. Meanwhile it is important to note that the statistical treatment given above indicates that Rankama's argument about *Corycium* is dependent on the primary carbon source of the biosphere differing from meteoritic carbon in the way in which it does in fact appear to differ.

II. THE METHANE CYCLE IN THE BIOSPHERE

Benedict and his coworkers have shown that fermentation in the alimentary tract of large herbivores gives rise to so much methane that a significant proportion of the carbon in the

	Methane per indiv. per day	Methane per indiv. per year	Population	Total Methane per year
Horse	106 gr.	37.7 kgm.	99.8 x 10^6	3862.10^3 metric tons
Cattle	200 "	73.0 "	510.9 x 10^6	37296.10^3 " "
Goat	14.7 "	5.4 "	116.8 x 10^6	631.10^3 " "
Sheep	15.1 "	5.5 "	532.2 x 10^6	2927.10^3 " "
Elephant	470.0 "	171.5 "	0.5* x 10^6	86.10^3 " "
				44802.10^3 " "

* Crude guess from ivory production of world.

organic matter ingested by these animals is returned to the atmosphere as CH_4 rather than CO_2. Mrs. Jane K. Setlow has prepared the preceding table, using the metabolic data of Ritzman and Benedict (1938) and the figures for approximate world populations collected by Rew (1929).

The total methane production of the biosphere must be very much greater than the estimated 45 million metric tons expired by these animals. Although it is possible that the estimates given above are based on animals better fed than the average, not only must extensive production by wild herbivores other than the elephant be considered, but also the vast methane production of lakes and swamps, for which a quantitative estimate is at present not possible. It would probably be quite safe to double the estimate derived from the table, giving a total annual production of 90.10^6 metric tons. Since the total photosynthetic fixation of carbon is about 150.10^9 tons per year (Riley 1944) it would appear that about 0.06% of the element circulating in the biosphere enters a subsidiary part of the carbon cycle as methane. It is also evident that this subsidiary path is truly cyclical, otherwise very large quantities of methane would have collected in the atmosphere.[1] At least two geochemical mechanisms of oxidation of this methane appear to exist, namely photochemical oxidation in the stratosphere and bacterial oxidation in soils. It is impossible to estimate which of these processes is likely to be quantitatively the more important.

It is well known that if an estimate be made of the total atmospheric oxygen together with the oxygen lost to the sediments in the oxidation of ferrous iron, the resulting quantity is very much less than the oxygen equivalent of the reduced carbon of the sediments. It seems, however, that such reduced carbon must have been mainly derived from CO_2 by photosynthesis. The excess carbon has been explained as derived from methane or carbon monoxide of volcanic origin, or as an apparent excess due to loss of gaseous oxygen through combustion of volcanic hydrogen. Since it seems likely that some hydrogen, carbon monoxide and methane

[1] A small stationary concentration would be expected and after this paper had gone to press my friend, Dr. Rupert Wildt, called my attention to the recent work of Marcel V. Migeotte (Astrophys. Jour. 107: 400-403, 1948) in which a qualitative demonstration of the presence of such a small stationary concentration in the earth's atmosphere is actually given.

would escape from volcanic vents prior to combustion, the demonstration of a large scale production and oxidation of methane in the biosphere is of some geochemical interest, particularly since the mechanisms responsible for oxidation of methane are likely also to operate in the oxidation of carbon monoxide and hydrogen. Conway (1942) has indicated the probable existence of similar mechanisms involved in the oxidation of hydrogen sulphide.

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A PRELIMINARY STUDY OF THE FLORA OF THE BRANDON LIGNITE

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ONE hundred years ago an unusually interesting and significant lignite deposit was accidentally discovered in the township of Brandon, Vermont. The deposit was encountered at considerable depth in the course of sinking a shaft for the extraction of underlying iron ore. The unexpected occurrence of brown coal in central Vermont almost immediately focused local scientific interest on the "Brandon lignite" and its geological and paleobotanical significance.

In 1851 Professor Shedd of the University of Vermont collected and sent specimens of the curious and abundant lignitized fruits from the deposit to Prof. Edward Hitchcock, then President of Amherst College and State Geologist of the Commonwealth of Massachusetts. Hitchcock visited the Brandon site the following spring and published a brief account of the Brandon lignite and associated strata which, he concluded, was "the type of a Tertiary formation hitherto unrecognized as such, extending from Canada to Alabama." (Hitchcock, 1853.) A collection of the seeds and fruits was sent by Hitchcock to Leo Lesquereux, the eminent Swiss-American paleobotanist. Lesquereux described and named a series of "species" of which several were botanically "identified" and the remainder assigned to the paleobotanical form-genus *Carpolithes* (Lesquereux, 1861).

The contributions of Hitchcock and Lesquereux, despite certain flaws in generalizations and conclusions, present a remarkably clear interpretation of the Brandon lignite deposit, particularly in view of the paucity of collections and data at their disposal. In retrospect their achievement is more apparent when it is recognized that little has been added to a satisfactory understanding of the deposit since the middle of the last century.

The Brandon deposit of lignite was accidentally "rediscovered" in 1901-02 in the course of clay mining operations at the same locality in which it was found in 1848. Rediscovery of the lignite coincided with the coal strike of 1902 and this combination of events led for a short time to the local utilization of the brown coal as a fuel. G. H. Perkins, State Geologist of Vermont, became keenly interested in the deposit, particularly its botanical aspects, and secured the coöperative interest of several paleobotanists and geologists in a study of the deposit and its geological relations. As a result of these investigations a series of papers were published on the flora of the Brandon lignites (Knowlton, 1902; Ami, 1906; Jeffrey and Chrysler, 1906; Perkins, 1904, 1906). Perkins' contributions, though extensive, were confined to descriptive analyses of the fossil fruits of which he enumerated 148 "species." Despite the completeness of the material he collected and examined, Perkins' studies provide little critical botanical information in the determination of the flora. As a result of his efforts, however, a very large series of the characteristic fossil forms from the lignite were adequately illustrated and carefully described as to gross features.

The geological relations of the deposit and associated strata were investigated and briefly described by T. Nelson Dale (1904) and J. B. Woodworth (1904). Their conclusions largely supplemented the earlier interpretation of the deposit made by Hitchcock.

The major problems centering on the lignite deposit at Brandon have remained unsolved despite the rather numerous but disjunct contributions during the last century. These problems are both paleobotanical and geological and comprise three major questions: (1) the botanical composition of the deposit and the significance of the flora in terms of phylogeny, vegetational history and paleoclimatology; (2) more exact determination of the age of the deposit and its possible correlation with other known Tertiary floras; (3) interpretation of the anomalous stratigraphic position of a presumably early or mid-Tertiary deposit occurring virtually at present stream level on a supposedly late Tertiary erosion surface.

Because of its unusual botanical and geological significance, the writers have undertaken an extensive investigation of the Brandon lignite. The summer of 1947 was spent in locating the lignite mass and in removing material for subsequent study. The lignite is concealed by both glacial drift and tailings from abandoned clay mining operations. Extensive core drilling was necessary to determine the position and depth of the deposit. Fortunately, exposure of a large mass of lignite was made within a few feet of the surface and small scale stripping and blasting operations revealed a considerable area of both lignite and associated organic sediments. Six tons of lignite were removed in the latter part of the summer and it is intended that additional material be obtained during the coming field season.

The Brandon deposit is located approximately 250 feet east of the Forestdale-Rutland road about one-half mile south of the village of Forestdale, Brandon. The lignite and associated clays, ochres, iron and manganese ores underlie a limited area on a tract of land owned by Mrs. Gertrude S. Horn of Forestdale. The deposit is close to the concealed and uncertain zone of contact of the Lower Cambrian Cheshire Quartzite and the Lower Cambrian Dunham Dolomite.

In contrast to many American lignites the Brandon deposit contains abundant remains of well preserved fruits and seeds as well as twigs, branches and larger wood fragments. The plant tissues are unmineralized and amenable to sectioning and anatomical study by the ordinary histological methods. In addition to fruits, seeds and woody fragments the deposit contains an abundance of diverse pollen grains, representing an estimated 100 species. Leaf remains are fairly common but are represented by small fragments only. With the exception of the pollen grains of *Pinus* all plant remains which have been thus far identified are those of dicotyledons and monocotyledons; the latter are represented by pollen grains only.

The presence of a wide range of various plant parts within a single deposit provides an opportunity to employ a highly desirable but little exploited paleobotanical "technique." This "technique" consists of utilizing concurrently all available collateral evidence in the botanical determination of the plants represented in the fossil flora. By this procedure determinations may be checked or verified with evidence derived from a wide range of sources, viz., wood structure, nodal anatomy, leaf epidermis, seeds, fruits and pollen. Correlation of the various lines of evidence makes it possible to verify or to preclude doubtful determinations based solely upon one category of evidence. The application of the method to the study of Tertiary floras necessitates extensive familiarity with the anatomy and morphology of living Angiosperms. In the opinion of the writers, however, it is the only feasible approach to the complex problems of identifying fossil Angiosperms.

Preliminary analysis of the Brandon flora shows the presence of characteristically warm temperate genera represented by *Nyssa, Cyrilla, Persea, Bumelia, Quercus* (live oak group), *Vitis* (section Muscadinia), *Symplocos*, and *Ilex*. With the exception of *Ilex*, all of the genera are represented by seeds, fruits, or wood. Tentative identifications of additional genera have been made on the basis of pollen grains. In view of the richness and complexity of the flora, however, collateral evidence is being sought to confirm many such identifications based upon pollen alone. Examination has been made of the pollen of over 400 living species of temperate, warm temperate, and sub-tropical species comprising a wide range of families of the dicotyledons. As a result of this survey the intrinsic difficulty of accurate identification of fossil pollen has been strongly impressed upon the writers. The occurrence of remarkably similar pollen types in remotely related groups of the Angiosperms clearly indicates the need for extreme care in the determination of genera. In contrast to the relatively depauperate woody floras of temperate Quaternary peats the pollen flora of Tertiary deposits present a far more complex picture. It is probable that the pre-glacial floras of northeastern America were richer in both species and genera than the forest associations of the same areas today. In the Brandon flora there is considerable evidence of several genera which are now confined to southeastern Asia. From the available evidence, however, the Brandon deposit represents an assemblage of genera now living in close association in the southeastern United States. In this connection it is significant that of 400 specimens of wood from the Brandon deposit a very large per cent represent the genus *Cyrilla*. *Cyrilla racemiflora* in the modern flora is a small tree or shrub abundant in the lowland swamp forests of the Gulf States.

One of the most interesting and significant groups represented in the Brandon lignite is the family Nyssaceae. As the result of histological studies a particularly striking series of fruits designated by Perkins as *Monocarpellites, Bicarpellites, Tricarpellites*, and *Glossocarpellites* have proved to be endocarps of Nyssaceous fruits. Several of the two and three carpellate forms are comparable to the London Clay fruits *Pro-*

tonyssa and *Paleonyssa* described by Reid and Chandler (1933). *Nyssa* endocarps of the "*Glossocarpellites*" series range up to 5 centimetres in length and average far larger than any fossil or living members of the *Nyssaceae* known to the authors. In the living flora the only approach to the larger Brandon forms is *Nyssa megacarpa*, a poorly known species from the mountains of Burma (Parker, 1929).

Nyssoid endocarps are remarkably abundant in the Brandon lignite. The genus *Nyssa* was apparently represented in the swamp forests of Brandon time by a larger number of species than exist today in eastern America. In addition, the numerous massive two and three carpellate endocarps unquestionably represent extinct genera closely allied to *Nyssa*. The nyssoid forms as a group are of exceptional interest in demonstrating a coherent phylogenetic series. The series clearly indicates that evolution in the *Nyssa* complex has involved both reduction in the number of carpels and reduction in thickness of the endocarp wall. The significant features of this phylogenetic series have been elucidated by histological study and anatomical comparison of the fruits of living *Nyssa* with the fossil forms.

The stratigraphic relations of the Brandon lignite deposit are complex and confused. In the absence of associated marine or fresh water invertebrates there is no direct method available for dating the deposit. The composition of the flora, its marked difference from the modern flora of similar latitudes, and particularly the presence of extinct genera all suggest considerable antiquity for the Brandon deposit. The obviously warm temperate aspect of the flora is strong evidence in favor of this view. From the accumulated and harmonious evidence regarding the Tertiary floras of western Europe, western and southeastern North America, it is apparent that the earlier Tertiary floras were characterized by the northward extension of many genera now limited to the lower middle latitudes of the north temperate regions. In view of these considerations it would seem logical to regard the Brandon flora as early or mid-Tertiary in age. Such an assumption, however, is based primarily upon paleoecological rather than direct floristic evidence or correlation.

The possible early or middle Tertiary age of the Brandon lignite is somewhat inconsistent with its present physiographic relationships. Because the lignite lies in the valley floor, almost at present stream level, it is difficult, without invoking special circumstances, to regard the deposit as older than Pliocene or at most late Miocene, the presumed age of the pre-glacial erosion surface on which the lignite was deposited. This interpretation is based on the assumption that western New England was subjected to a series of uplifts in early and middle Tertiary time with the resulting formation of erosion surfaces discernible in the present topography. If this view of the physiographic history be correct it is obviously difficult to assign the lignite deposit to the early Tertiary, the age which is more consistent with paleobotanical evidence. On the other hand there is no critical botanical evidence to establish the Eocene age of the Brandon flora as proposed by Berry and others (Berry, 1919). It is apparent, therefore, that the problem of the age of the Brandon lignite cannot be resolved until more is known of its geological relations and its possible paleobotanical correlation. It should be emphasized, however, that the deposit and its associated clays, ochres, and ores cannot be referred to the early Tertiary without necessitating a thorough reinterpretation of the Cenozoic erosional history of western New England.

It is proposed to summarize the results of present studies of the Brandon lignite in a series of forthcoming papers dealing with discrete phases of the problem. Tertiary floras of the southern and western United States have been intensively studied within the last few decades (Chaney, 1940). Similarly numerous critical studies have recently been made of the Cenozoic deposits of western and central Europe (Kirchheimer, 1939, 1942; Thiergart, 1940; Hirmer, 1942). In marked contrast, the composition of the Tertiary flora of the northeastern United States is virtually unknown, largely because of its meagre preservation in the fossil record. The Brandon lignite deposit is one of an exceedingly few clues to the origin and development of the pre-glacial vegetation of northeastern North America.

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THE GALLIUM CONTENT OF IGNEOUS ROCKS

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ABSTRACT. Analysis of composite samples indicates a Ga/Al weight ratio varying from 1.5×10^{-4}, or somewhat less, in basic rocks to 3×10^{-4} in granites high in silica. There is some indication that gallium may replace ferric iron in minerals. The abundance of gallium in the upper lithosphere is estimated to be 15 parts per million.

THE most comprehensive investigation of the geochemistry of gallium was made by Goldschmidt and Peters (1), who particularly pointed out the close association of gallium and aluminum in silicates and showed that the ratio of gallium to aluminum increased in late differentiates. In a later paper, Goldschmidt and coworkers (2) reported figures for the gallium content of German granites, gabbros, shales, and greisen. Since that time, data on the abundance of gallium in rocks have been given by van Tongeren (3), Rankama (4), Wager and Mitchell (5), and Sahama (6, 7), as well as others. Nockolds and Mitchell, in a noteworthy contribution (8), have reported the gallium content of the component minerals of Caledonian plutonic rocks from western Scotland, as well as of the rocks themselves. Erämetsä (9) studied the distribution of gallium in Finnish minerals, mostly silicates. Other workers, especially Russian, have also investigated the occurrence of gallium in minerals.

In the present work an attempt has been made to fix somewhat more exactly than previously, the gallium-aluminum ratio in igneous rocks and at the same time to obtain some data on the abundance of the element in common rocks. For this purpose, samples previously analyzed for the major constituents were taken, and determinations were run on composite samples (Table I) and also on some individual samples (Table II).

METHOD OF ANALYSIS

Previous workers have determined trace concentrations of gallium by optical spectrography. The present results were obtained by an independent method which involves extraction of gallium chloride by ethyl ether from a hydrochloric acid solution of the decomposed sample, followed by fluorimetric determination of gallium as the 8-hydroxyquinolate in chloro-

TABLE I

Gallium in Composites of Igneous Rocks

Silica Range % SiO_2	Average Silica % SiO_2	Average Alumina % Al_2O_3	Rock Type and Number of Samples	Region	Gallium Content P.p.m. Ga	Ga/Al x 10^4
80.6–76.3	77.7	11.6	Granite (12)	U. S.	18.5 16.5 } 17.5	2.9
81.5–75.1	77.0	11.7	Granite (8)	U. S. and Canada	18.5	3.0
77.4–71.5	74.1	13.75	Granite, etc. (9)[a]	California	15	2.1
76.8–68.2	73.1	13.7	Granite (5)	Texas (Llano region)	17.5 17.0 } 17	2.3
74.3–70.6	72.6	14.2	Granite (7)	Canada	16.5	2.2
73.9–70.0	72.0	14.5	Granite (9)	Africa	14.5 16 } 15	2.1
69.6–65.5	68.4	15.3	Granite, etc. (14)[b]	c	15.5	1.9
65.7–52.7	62.0	15.9	Diorite (7)	Oregon and California	15	1.8
57.7–44.4	50.0	15.7	Basic (24)[d]	e	13.5 11 } 12 11.5	1.4
51.3–45.7	47.6	18.8	Gabbro-diorite (4)	California	17	1.7
50.4–46.0	47.4	16.8	Basalt	Michigan[f]	17.5	2.0
54.7–33.4	42.8	1.7	Sub-basic (6)[g]	h	1.5	1.5±
53.2–48.3	50.7	28.7	Anorthosite (3)	New York, Minnesota and Montana	20[i]	1.3
64.1–56.0	60.3	17.5	Syenite (6)	New York and Minnesota (1)	21 22 } 22	2.4
57.9–49.6	53.1	18.3	Phonolite (4)	Africa	22	2.3

[a] 4 granites, 2 quartz monzonites, 1 monzonite, 2 granodorites.

[b] 6 granites, 1 granite porphyry, 1 feldsite porphyry, 1 porphyry, 1 quartz monzonite, 1 granodiorite, 1 diorite, 2 dacites.

[c] Canada 2, New York 1, Texas 2, Montana 2, Wyoming 1, California 2, 4 from unspecified U. S. or Canadian localities.

[d] 9 gabbros, 1 norite, 4 diabases, 8 basalts, 1 peridotite, 1 pyroxenite.

[e] Eastern and Western U. S. 13, Newfoundland 1, Africa 1, Antarctica 3, India 1, unspecified localities (probably U. S.) 5.

[f] Weighted composite of Greenstone flow.

[g] 1 peridotite, 1 dunite, 1 harzburgite, 1 bronzitite, 1 pyroxenite.

[h] Newfoundland 3, U. S. 2.

[i] Average from 3 individual anorthosite values in Table II.

form solution (10). Ordinarily a 0.2 or 0.25-g. sample is taken and a 4/10 aliquot of the solution is used for the determination. As little as 1 part per million of gallium can be detected in the sample. If the amount of sample is limited, the analysis can be made on 0.05 g. of material. The fluorimetric determination is made visually by comparison against a standard. The precision of the comparison corresponds to an error of about 10 per cent. The recovery of known amounts of gallium carried through the procedure averaged better than 90 per cent for the range 5 to 50 p.p.m. of gallium. To compensate the tendency toward slightly low results, the experimental values have been multiplied by the factor 1.1, and these corrected values are the ones given in the tables.

TABLE II

Gallium in Various Igneous Rocks

	Al_2O_3 %	Fe_2O_3 %	Ga P.p.m.	Ga/Al x 10^4
Rhyolite, Cook County, Minn. (12, Table 3, No. 1)	13[a]	—	16	2.5±
Pantellerite	8.9	7.3	41	8.7
Diorite, Little Falls, Minn. (12, Table 3, No. 17)	17.5	3.7	16.5	1.8
Diabase, Grand Marais, Minn. (12, Table 3, No. 9)	14.5	7.4	19	2.4
Diabase, Lake County, Minn. (12, Table 3, No. 14)	21.0	3.3	15.5	1.4
Harzburgite	5.0	4.0	4	1.5
Peridotite	3.0	0.7	2	1+
Dunite	0.4	0.8	<1	—
Augite syenite	15.1	6.0	25	3.1
Shonkinite	8.5	7.7	23	5
Anorthosite, Montana	28.9	0.6	19	1.3
Anorthosite, Split Rock, Minn.	30.7	0.2	20	1.2
Anorthosite, New York State	26.6	0.4	21	1.5

[a] Approximate value; figure from an old analysis of a different specimen from the same locality.

DISCUSSION OF RESULTS

The gallium-aluminum ratio. On the assumption that the distribution of gallium is controlled chiefly by aluminum, weight ratios of gallium to aluminum have been calculated for the samples listed in Tables I and II. The variations of Al_2O_3 and of the ratio Ga/Al with silica content is shown in Figure 1. It is seen that Ga/Al generally increases as differentiation proceeds toward silicic or alkali-rich products. This ratio has a value of 1.5×10^{-4} or slightly less for the more basic rocks and for anorthosites, and increases to 3×10^{-4} in the high-silica granites. Between these two average extremes fall the values for intermediate rocks and the less silicic granites. The ratio is uncertain for rocks very low in aluminum because the error in the determination of the very small amount of gallium present is then large, but it seems to be near 1 or 1.5×10^{-4}.

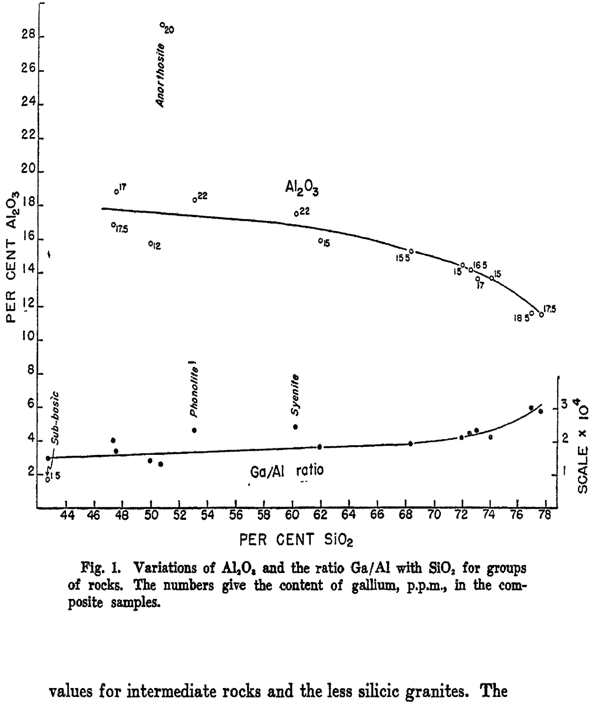

Fig. 1. Variations of Al_2O_3 and the ratio Ga/Al with SiO_2 for groups of rocks. The numbers give the content of gallium, p.p.m., in the composite samples.

For the granites and anorthosites, in which most of the gallium is present in the feldspars, the ratios given should be close to the substitution ratios for gallium in the lattice. When relatively much ferric iron is present in a mineral there would seem to be a possibility that gallium can substitute to some extent for iron as well as for aluminum. Goldschmidt's empirical ionic radii for aluminum, gallium, and ferric iron are respectively 0.57, 0.62, and 0.67A. Apparently the ionic sizes of trivalent iron and gallium are sufficiently similar to permit mixed crystal formation. Because of the narrow range over which Ga/Al normally varies, the relatively large possible error in the determination of gallium, the general preponderance of aluminum over ferric iron in igneous rocks and the possibility of oxidation of ferrous iron to ferric, it is difficult to demonstrate convincingly the replacement of ferric iron by gallium. However, in some rocks of the basic type, in which no particular enrichment of gallium is apparent, the gallium-aluminum ratio is significantly greater than in silicic rocks. Thus, a diabase from Grand Marais, Minnesota (Table II), shows a Ga/Al ratio of 2.4×10^{-4} (from the mean of two concordant gallium determinations), which is equal to, or greater than, that of ordinary granites. This rock has a high content of ferric iron (7.4 per cent Fe_2O_3).

Erämetsä (9) considered the possibility of replacement of ferric iron by gallium, but was unable to test the hypothesis on his samples spectrographically because of line coincidences between the two metals when much iron was present. Tsarovs'kiĭ (11) in studying the gallium content of the minerals of the nepheline pegmatites of the Azov region found the amount of gallium to increase with aluminum in the light-colored minerals and with the sum of aluminum and iron in the dark colored ones.

The ionic radii of trivalent chromium and quadrivalent titanium (each 0.64 A, according to Goldschmidt) are so close to the radius of gallium that the entrance of gallium into the minerals of these elements must be considered a possibility. Goldschmidt and Peters (1) found 5 to 10 p.p.m. of gallium in chromites and 10 p.p.m. in two specimens of magnetite-ilmenite. The aluminum content of these samples is not stated, and it is not known whether gallium shows any enrichment compared to aluminum in them. In ordinary igneous rocks these possible substitutions are of minor quantitative importance. Nockolds and Mitchell (8) found a little gallium in early titanites, but none in the later ones.

The tendency for gallium to be concentrated in alkalic rocks, as pointed out by Goldschmidt for nephelites, is illustrated by the high gallium content of the pantellerite of Table II. This rock has a gallium-aluminum ratio of almost 10^{-3}. Even the $Ga/(Al + Fe^{III})$ ratio is abnormally high. It seems likely that both gallium and ferric iron have been enriched compared to aluminum because of their greater ionic radii.

Nockolds and Mitchell found the ratio Ga/Al in their rocks (mostly gabbros, diorites and granodiorites) to have the approximate value 2×10^{-4}. Their results are thus much the same as ours, although they report that the ratio remains practically constant throughout the series. However, these authors found Ga/Al to be greater in late than in early hornblendes and biotites. The highest values for the ratio in hornblendes were found in two specimens separated from a tonalite and a granodiorite, which yielded the figures 7×10^{-4} and 8×10^{-4} respectively. In the present work, an alkalic amphibole separated from a New Hampshire granite (12, Table 9), of which it makes up 5 per cent, was found to contain 19 p.p.m of gallium. The Al_2O_3 and Fe_2O_3 content of this amphibole are respectively 2.5 and 7.5 per cent. The ratio Ga/Al is 1.4×10^{-3}. The granite itself was not available for analysis, but it seems unlikely that its Ga/Al ratio would be much greater than 3×10^{-4}, the average for high silica granites.

Nockolds and Mitchell observed that the Ga/Al ratio is lower in early plagioclases than in the later. Their values of the ratio for three plagioclases separated from two gabbros and an appinitic diorite respectively (An = 63, 59, and 45) are slightly greater than 1×10^{-4}, and thus are the same, within the experimental error, as the values obtained for anorthosites (Table II) in the present analyses. In plagioclases of more acid rocks, these authors found Ga/Al to average 3×10^{-4}. It is interesting to note that the ratio is almost uniformly lower (average about 2×10^{-4}) in the potassium feldspars than in plagioclases. Shimer (13) found an average of 40 p.p.m. (from 10 to 60 p.p.m.) of gallium in feldspars separated from New England granites. This content appears to correspond to a Ga/Al ratio somewhat greater than reported by Nockolds and Mitchell for feldspars of the Caledonian rocks.

Abundance of gallium in the lithosphere. The rather monotonous uniformity in the gallium content of most igneous rocks simplifies the problem of obtaining an average value for gallium in the lithosphere. From the figures in Table I an average of 17 p.p.m. of gallium in granites may be derived. For the common basic rocks a content of 12 p.p.m. may be taken, this average derived from a composite of 24 samples being considered more representative than the other values for basic rocks in the table, although it may be a trifle low. On the assumption that equal weights are to be given to the common basic rocks and to granites in the upper lithosphere, and that the gallium content of intermediate rocks falls between that of these two groups, an average value of 15 p.p.m. is obtained for the gallium content of the earth's crust.

Another method of calculation may be based on the gallium-aluminum ratio of rocks. Taking 2.4×10^{-4} as the average ratio for granites and 1.4×10^{-4} as the average for basic rocks, we obtain a mean of 1.9×10^{-4} which also may be taken to represent the ratio for intermediate rocks. Accepting Clarke's value of 8.1 per cent for the abundance of aluminum in the lithosphere, we arrive at the same figure as before, 15 p.p.m. of gallium. Too much significance should not be attached to this exact value, but it seems unlikely that the abundance is less than 10 p.p.m. or greater than 20 p.p.m. On the basis of 15 p.p.m. of gallium, the atomic ratio of gallium to aluminum in the crusts is 7.5×10^{-5}.

Comparison with previous results. The earlier spectrographic determinations of gallium in silicate rocks and minerals were for the most part semi-quantitative and close agreement with the present results is not to be expected. If this limitation is taken into consideration, our results are substantially the same as those obtained by Goldschmidt and coworkers. Goldschmidt and Peters (1) estimated the average gallium content of granite to be approximately 20 p.p.m. (30 p.p.m. Ga_2O_3), and that of basic plagioclase rocks to be a little less than 10 p.p.m. (10 p.p.m. Ga_2O_3). In a later paper (2) the average 15 p.p.m. of gallium is reported for 14 German granites and slightly less than 10 p.p.m. (10 p.p.m. Ga_2O_3) for 11 German gabbros. For the abundance of gallium in the lithosphere, Goldschmidt (14) arrived at the value 15 p.p.m., the same as calculated above.

Van Tongeren's average for gallium in 49 Dutch East Indies granites of Mesozoic age is 50 p.p.m., with individual values extending from 8 to 375 p.p.m., but mostly in the range

20 to 75 p.p.m. (3, quoted in 15). For gabbros from this region, an average of approximately 40 p.p.m. gallium (50 p.p.m. Ga_2O_3) was found. These amounts are much higher than those found here for American rocks. Rankama (15) is of the opinion that there may be a difference in the gallium contents of pre-Cambrian and younger granites.

In a paper on the trace elements of the rocks of southern Finnish Lapland, Sahama (6) reports an average of approximately 75 p.p.m. gallium for granites and 20-25 p.p.m. (30 p.p.m. Ga_2O_3) for gabbros and dolerites. These spectrographic figures are based on the visual comparison of line intensities and Sahama points out that they should be regarded as giving the order of magnitude and that the results may be in error by the factor 2 or 3. Moreover, he states that further work is likely to lower these values, rather than to raise them. In a standard mixture of East Fennoscandian rapakivi granites, Sahama found approximately 75 p.p.m. of gallium (7).

Rankama (4) obtained results very similar to those of Sahama for gallium, in an earlier study of other Finnish rocks.

Wager and Mitchell (5) give 25 p.p.m. as the gallium content of the olivine-gabbro magma of the Skaergaard intrusion, East Greenland. This value is based on the analysis of the chilled facies of the marginal contact. The gallium concentration of the various rocks of the layered differentiation series was found to vary from 10 to 20 p.p.m., in sympathy with the aluminum content. No trend in the value of the gallium-aluminum ratio is evident, the average value being slightly greater than 2×10^{-4}. Gallium was evidently determined to the nearest 5 p.p.m. These values are not greatly different from those for corresponding basic rocks analyzed here, but they tend to be higher.

In the recent work of Nockolds and Mitchell (8), to which reference has already been made, the gallium content of rocks of western Scotland (mostly gabbros, diorites, and granodiorites) was found to run from 15 to 20 p.p.m., very rarely higher or lower. These results agree with corresponding values in Table I.

On the whole, one obtains the impression of great uniformity in the gallium content of the common igneous rocks, independent of age and locality.

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DENDRASTER ELSMERENSIS DURHAM, N. SP.[1]

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ABSTRACT. The new species, *Dendraster elsmerensis* Durham, with a marginal periproct is described. The plate systems of the oral surface, the internal structure, and the pattern of the ambulacral food grooves of the genera *Anorthoscutum*, *Dendraster*, *Echinarachnius* and *Merriamaster* are described and compared. It is concluded that the four genera are distinct, with *Anorthoscutum* being the most primitive, and the genotype of *Echinarachnius* somewhat advanced, but still more primitive than *Dendraster* and *Merriamaster*. However, the middle Miocene species *E. merriami* is more primitive than any of the other genera and could possibly represent the ancestral stock.

THE occurrence of an echinoid which closely resembles *Dendraster excentricus* (Eschscholtz) in the basal beds of the Pico formation in Elsmere Canyon, near Newhall, California, has long been known (Pack, 1909, p. 282; English, 1914, p. 209; Kew, 1920, p. 122; Woodring, Stewart and Richards, 1940, p. 111). All workers who have recorded their observations have identified the specimens from this locality as *D. excentricus*, and this occurrence of the living species has been cited (Kew, 1924, pp. 79-80) either as indicating a long time range for *D. excentricus*, or else as indicating a later age for the containing beds than lower Pliocene.

The original echinoid material upon which Pack, English, and Kew based their statements is in the University of California collections, locality numbers 1601 and 1602.[2] Examination of the specimens from these localities shows that they rather closely resemble *D. excentricus*, but are all fragmentary and none show the periproct. Recently additional specimens (loc. A 4457) were collected from what is assumed to be near the same locality.[3] and certainly is at the same horizon as the original material.

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[1] A contribution from the Museum of Paleontology of the University of California, Berkeley, California.

[2] All locality and specimen numbers used in this paper refer to the Invertebrate Collections, Museum of Paleontology, University of California, Berkeley, California.

[3] Loc. A 4457 — apparently in SW ¼ of NE ¼ of Sec. 8, T 3 N, R 15 W, San Bernardino Base and Meridian. From hillside on northwest side of stream, about 200 yards upstream from point where contact of Pico formation and basement cross the stream. Collected from 0′ to 25′ above base of Pico formation. In association with numerous molluscs and *Astrodapsis fernandoensis* Pack.

Some of the specimens are complete and moderately well preserved. On all individuals (5) in which that part of the test is preserved, it can clearly be seen that the periproct is marginal, a feature which immediately removes this species from the genus *Dendraster*, according to the conventional criteria used by most present day American workers in this field (as exemplified by Grant and Hertlein, 1938, pp. 53, 78). However, in all other respects, the specimens are typically ***Dendraster***, and it is worthy of note that Stewart (Woodring, Stewart and Richards, 1940, p. 81) has shown that in *D. coalingensis* Twitchell, the periproct is occasionally marginal or supramarginal. Because of the above anomaly, a review of the allied genera *Echinarachnius*, *Anorthoscutum* and *Merriamaster* was undertaken during the description of the species and it was concluded that the Elsmere Canyon species is a primitive member of the genus *Dendraster*. The description of the new species is as follows:

Genus *Dendraster* Agassiz

Dendraster Agassiz, in Aggasiz and Desor (1847), Ann. Sci. Nat. (Zool.), ser. 3, vol. 7, p. 135; Grant and Hertlein (1938), Univ. Calif. Los Angeles, Publ. Math. Phys. Sci., vol. 2, pp. 78-79.

Dendraster elsmerensis Durham, n. sp.

Pl. 1, figs. 2, 3, 4, 6; Text-fig. 2F

Echinarachnius excentricus Esch., English (1914), Univ. Calif. Publ., Bull. Dept. Geol. Sci., vol. 8, p. 209.

Dendraster excentricus (Eschscholtz), Kew (1920), Univ. Calif. Publ., Bull. Dept. Geol. Sci., vol. 12, pp. 121-122, *partim* (lower Pliocene occurrence only).

Dendraster excentricus Eschscholtz, Grant and Hertlein (1938), Univ. Calif. Los Angeles, Publ. Math. and Phys. Sci., vol. 2, pp. 84-87, *partim* (lower Pliocene occurrence only).

Test moderately small, margins thin, apical system moderately excentric, periproct marginal; in general appearance quite similar to *D. excentricus* except for its smaller size and marginal periproct; center of apical system about two-thirds the length of test from anterior margin; test variable in outline, from elongate ovate to shortened ovate; length of right petal of trivium slightly over two-thirds the radius of test from center of apical system (three specimens), as compared to slightly under two-thirds for the corresponding radius of *excentricus*; length of anterior petal of trivium between two thirds and three-fourths of the corresponding radius, as compared to about three-fifths for *excentricus*; angle between the bivium less than in *excentricus* or *gibbsii*, and about the same as in *vizcainoensis*; peristome 0.42 to 0.45 length of test from posterior margin; about ten tubercles across middle of anterior petal between pores, compared to about sixteen in a similar position for *excentricus*, and four or five for a new species from the Pliocene of the Gulf of California area; internal supporting walls and pillars of test fewer than in *excentricus*, being about comparable to those of *Echinarachnius parma*; primordial ambulacral and interambulacral plates both present in basicoronal row, both moderately small, but with interambulacral plates larger than ambulacral (text fig. 2F), only ambulacral plates present in penultimate row; auricles slightly oral from center of interambulacral plates, character not specifically determinable but apparently similar to *excentricus*; pattern of ambulacral and interambulacral plates on oral surface generally similar to that of *excentricus* except that there are fewer ambulacral plates between the ends of the petals and the ambitus, whereas the number of interambulacral plates is approximately the same.

Dimensions: holotype no. 32949, length 43 mm., width (incomplete) 44 mm., height 7.3 mm.; paratype no. 32950, length 34 mm., width (incomplete) 33 mm., height 5.5 mm.; paratype no. 32956, length (incomplete, estimated) 37 mm., width (incomplete, estimated) 39 mm.

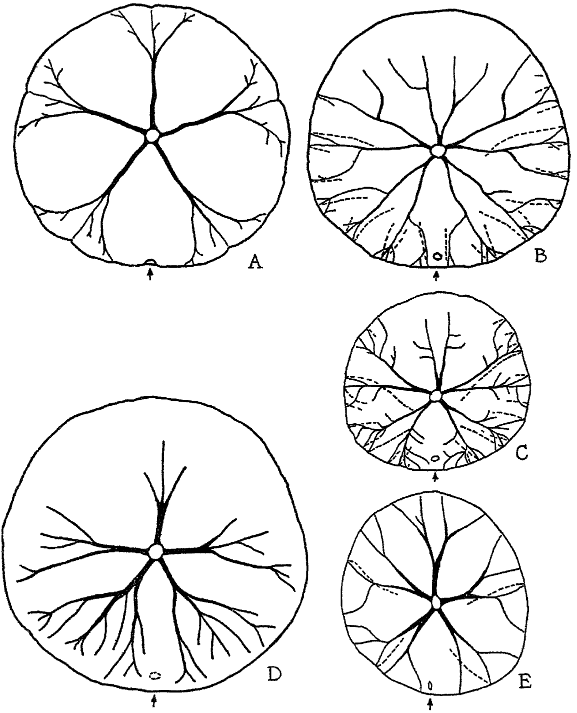
Types: holotype no. 32949; paratypes nos. 32950, 32951, 32956, 32957; all specimens from loc. A 4457; paratype 32953 from loc. 1602. Two paratypes deposited in Paleontological collections of the California Academy of Sciences, San Francisco.

Age: The associated fauna has been commonly considered as of lower Pliocene age.

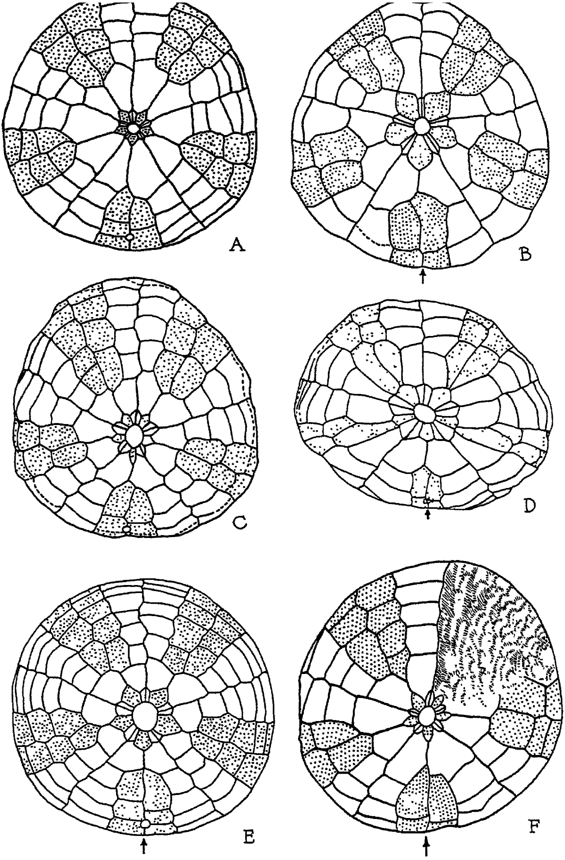
Discussion.—The marginal periproct separates this species from all known species of *Dendraster*. Superficially except for the position of the periproct it is quite similar to *D. excentricus*. However, the fewer number of tubercles across the width of the anterior petal readily separate the two, and the lesser angle between the bivium is also characteristic. Internally the species is readily separable from *D. excentricus* by the relatively few internal supporting walls and pillars—being closely similar to *Echinarachnius parma* (pl. 1, fig. 8) in abundance and pattern. Possibly it may have even fewer concentric pillars than that species. Sections of *D. excentricus* (pl. 2, fig. 4), *D. gibbsii* (Remond) (pl. 2, fig. 2), *D. coalingensis* Twitchell (pl. 2, fig. 1) and a new species from the Pliocene of the Gulf of California area have been made and compared with the present species. All except *D. gibbsii* are closely similar to *excentricus* rather than to *elsmerensis*, possibly an indication of a primitive trait in the last named form. *D. gibbsii* differs from all other species investigated in the markedly heavy character of the partitions and pillars. The pattern of the food grooves on the oral surface of *elsmerensis* is not readily discernible, but on paratype no. 32953, part of it can be faintly seen and appears to be similar to that of *D. excentricus*.

The non-conformity of the periproct of *D. elsmerensis* to the conventional criteria of the genus *Dendraster* led to an investigation of the pattern of the ambulacral grooves, the internal structure, and the plate systems of the Pacific Coast genera *Anorthoscutum*, *Dendraster*, *Echinarachnius* and *Merriamaster* as members of an apparently closely related group. Specimens of the genotypes and other species of each of the genera except *Anorthoscutum* were sectioned and examined by various means. No well-preserved specimens of the type of *Anorthoscutum* (*Scutella interlineatum* Stimpson) were available for preparation, but specimens of *A. oregonense major* and *A. oregonense quaylei*, both closely related to the type, were utilized. Available specimens of *Anorthoscutum interlineatum* were adequate to show that they have the same arrangement of the plates on the oral side of the test as *A. oregonense quaylei*. Because all other characters are similar, it has been assumed that the pattern of the ambulacral grooves is basically the same as in *A. oregonense major*, the species of this genus in which they have been observed.

Text figure 1 illustrates the pattern of the ambulacral grooves on the oral surface of the above genera. It is readily seen that there are four principal types—corresponding to the four genera, with that of *Echinarachnius* being markedly different from the rest in that each ambulacral unit is nearly symmetrical with the others. The pattern of *Echinarachnius parma* also differs from that of the other genera in that the



Text fig. 1. Pattern of ambulacral grooves on oral surface (dotted lines indicate continuation on aboral surface): A, *Echinarachnius parma* Lamarck) (XO.7), hypotype no. 32933 (loc. A 4455); B, *Dendraster excentricus* (*Eschscholtz*) (XO.7), hypotype no. 32934 (loc. A 3783); C, *Dendraster gibbsii* (Remond) (XO. 7), hypotype no. 32940 (loc. A 4321); D, *Anorthoscutum oregonense major* (Kew) (XO.7), cotype no. 11352 (loc. 71); E, *Merriamaster perrini* (Weaver) (XO. 7), hypotype no. 11062 ("Coalinga District"). (Arrow indicates periproct on all figures.)



Text fig. 2. A, *Dendraster excentricus* (Eschscholtz) (X0.75), hypotype no. 32958 (loc. A 3783). Plate system on oral surface (interambulacral plates stippled); B, *Echinarachnius parma* (Lamarck) (X 1), hypotype no. 32931 (loc. A 4456). Plate system on oral surface (interambulacral grooves do not branch until over halfway to the margin where the main groove gives off two lateral branches of approximately equal size, then all three continue to the margin with a few short minor branches on each main branch.

The genus *Merriamaster* as represented by the genotype, *M. perrini*, also appears to have nearly symmetrical branching on the oral surface, and is usually considered to have the grooves restricted to the oral surface (Grant and Hertlein, 1938, p. 53, key), and on most specimens such is apparently the case. However, a single large (length 51.5 mm.) well-preserved specimen (text fig. 1-E) (hypotype no. 11062, "Coalinga District") clearly shows faint extensions of the grooves onto the oral surface in all except the anterior ambulacrum. This same specimen also shows secondary branching on the oral surface near the margin, in contrast to the simple furrows shown by Stewart (Woodring, Stewart and Richards, 1940, pl. 46, figs. 1, 2). However, his specimens are much smaller than the one here noted, and inasmuch as all other specimens available to the present author (mostly of the same size as Stewart's) show the same type branching as his, it may be possible that the more complex branching does not begin until large size is attained (and can only be observed on well-preserved specimens). The extension of the grooves onto the oral surface is a feature in common with *Dendraster*, but the entire system is much less complex than that of that genus (see text fig. 1-B, C). A further difference between the two genera is in the greater development of the grooves of the anterior ambulacrum of *Merriamaster*.

The ambulacral grooves of the genus *Anorthoscutum* as typified by those of *A. oregonense major* (text fig. 1-D) are somewhat similar to those of *Dendraster* in that the system of the two posterior ambulacra is the most complex, while that of the anterior ambulacrum is the most reduced. However,

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plates stippled, arrow indicates periproct); C, *Dendraster gibbsii* (Remond) (X1.5), hypotype no. 32941 (loc. A 4321). Plate system on oral surface (interambulacral plates stippled); D, *Anorthoscutum oregonense quaylei* Grant and Hertlein (X1.5), topotype no. 32943 (loc. 1881). Plate system on oral surface (interambulacral plates stippled, arrow indicates periproct); E, *Merriamaster perrini* (Weaver) (X1.9), hypotype no. 32948 (loc. A 4320). Plate system on oral surface (interambulacral plates stippled); F, *Dendraster elsmerensis* Durham, n. sp. (X1.5), paratype no. 32950 (loc. A 4457). Plate system on oral surface (interambulacral plates stippled, arrow indicates position of periproct).

there is no evidence of any extensions on the aboral surface, and the branching is usually a modified trifurcation instead of a bifurcation. Further, the main branching takes place at a point about one-third out from the peristome instead of very close to it.

The basic pattern of the ambulacral grooves appears to be much the same in all species of *Dendraster* examined, with the greatest differences being found in *D. gibbsii* (text fig. 1-C) as contrasted with *D. excentricus* (text fig. 1-B). As in *Anorthoscutum*, the system of the anterior ambulacrum is the simplest, while those of the two posterior areas extend over the margin onto the aboral surface. Generally each ambulacral groove bifurcates once close to the peristome, and a second time, somewhat unequally, about a half or two-thirds the distance to the margin, and subsequently several short minor branches are given off. *D. gibbsii* differs from *D. excentricus* in that the minor branches are more numerous and that the second bifurcation does not differ greatly in prominence from the subsequent minor branches.

In the ambulacral and interambulacral plate systems (text fig. 2) of the oral side of the test, marked differences are apparent between some of these genera and close similarities between others. The arrangement of the plates in *Merriamaster* and *Dendraster* is quite similar, but *Echinarachnius* and *Anorthoscutum* are markedly different from each other and from the other two. The plates of the basicoronal row surrounding the peristome in each case consists of ten ambulacrals and five interambulacrals, thus conforming to the typical clypeastroid pattern of the suborder (Jackson, 1912, pp. 71, 72, text fig. 52). Comparatively speaking, the plates of the basicoronal row

PLATE 1.

1. *Anorthoscutum oregonense quaylei* Grant and Hertlein (X1.4), topotype no. 32943 (loc. 1881). Oral surface.

2, 3, 4, 6. *Dendraster elsmerensis* Durham, n. sp. 2.—(X1.2), paratype no. 32957 (loc. A 4457), natural section, internal view of plates surrounding mouth; 3.—(X1.1), aboral view of paratype no. 32956 (loc. A 4457); 4.—(X1.25), holotype no. 32948 (loc. A 4457); 6.—(X1.25), paratype no. 32950 (loc. A 4457)—note tubercles in anterior ambulacrum.

5. *Anorthoscutum oregonense quaylei* Grant and Hertlein (X1.4), holotype no. 32944 (loc. 1883). Partial section to show internal partitions.

7. *Anorthoscutum* sp. (X1.25), hypotype no. 32942 (loc. A 3779). Partial section of a small specimen to show internal partitions and position of auricles.

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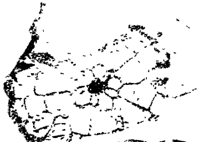

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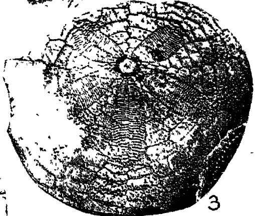

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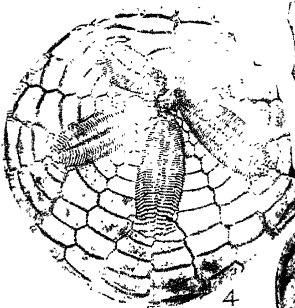

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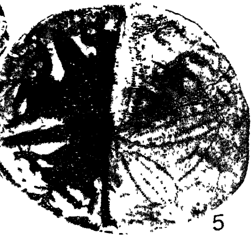

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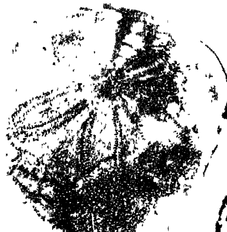

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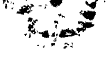

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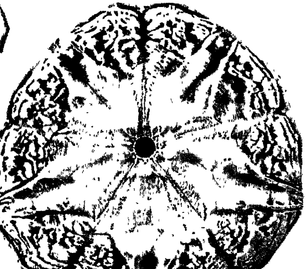

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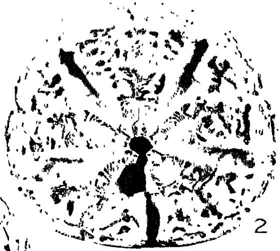

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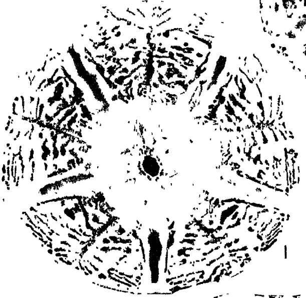

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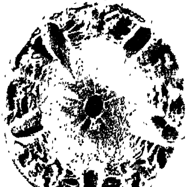

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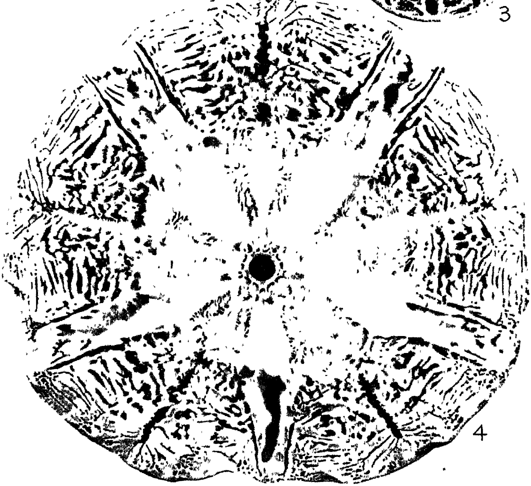

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are smallest in *Dendraster*, and successively larger in *Merriamaster*, *Anorthoscutum*, and *Echinarachnius*.

On the basis of the arrangement of the plates, *Anorthoscutum* (text fig. 2-D) appears to be the most primitive genus in that the interambulacral plates of the basicoronal row are in contact with subsequent plates of the interambulacrum in areas 1, 2, 3 and 4, and only in area 5 have the interambulacral plates been pushed dorsally so that the first post-primordial ambulacral plates of areas I and V are in contact. In interambulacral areas 1 and 4, only one column of plates is in contact with the primordial plate, with the plate making this contact being very high and rather narrow. In areas 2 and 3, both columns make contact with the basicoronal plate. The plates of the basicoronal row are moderately large and the interambulacral plates in areas 2 and 3 are larger than the other three. On the oral surface, there are from three to six post basicoronal rows of plates in each ambulacral column. In the interambulacral areas, there are two post basicoronal plates to a column in area 5, two in the anterior and three in the posterior columns of areas 1 and 4, two in the posterior and three in the anterior columns of areas 2 and 3. The interambulacral plates of the basicoronal row are from one to three times as large as the adjacent ambulacrals.

In *Echinarachnius* (text fig. 2-B), in contrast to *Anorthoscutum*, the interambulacral plates of the basicoronal row are just barely in contact in areas 2 and 3 with the succeeding plate of the columns and are completely separated from them in areas 1, 4 and 5. The primordial interambulacral plates of the basicoronal row have become greatly enlarged in contrast to the adjacent primordial ambulacrals and have a total area six to eight times as great. There are three to four post-basicoronal plates in each ambulacral column on the oral surface, but only two interambulacral plates to a column. The

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PLATE 2

1. *Dendraster coalingensis* Twitchell (X1.25), hypotype no. 32938 (loc. A 3346). Internal view of sectioned specimen.
2. *Dendraster gibbsii* (Remond) (X1.25), hypotype no. 32939 (loc. A 4321). Internal view of sectioned specimen.
3. *Merriamaster perrini* (Weaver) (X1.25), hypotype no. 32947 (loc. A 4320). Internal view of sectioned specimen.
4. *Dendraster excentricus* (Eschscholtz) (X1.25), hypotype no. 32935 (California). Internal view of sectioned specimen.

shape of the primordial interambulacral plates is considerably different than that shown by Jackson (1912, p. 80, fig. 52) for a specimen of the same species from Eastport, Maine. The specimen here studied is from Kodiak Island, Alaska. The marked differences between the shape of these particular plates indicate that specimens of this species (*Echinarachnius parma*) from all parts of its range should be carefully studied to see if there may not be some geographic differentiation present.

Dendraster (text fig. 2-A, D, F), *Merriamaster* (text fig. 2-E), and *Astrodapsis* (not figured) are characterized by the plates of the basicoronal row being quite small in comparison to the next row of plates, and by all the interambulacral areas being approximately equally separated from the basicoronal interambulacral plates, so that the first row of post-basicoronal ambulacral plates of each ambulacrum is in lateral contact with the plates of the adjacent ambulacrum. In addition, in contrast to the condition prevailing in *Echinarachnius* but similar to that in *Anorthoscutum*, the first post-basicoronal ambulacral plates are considerably larger than subsequent plates of the same column. In *Dendraster excentricus* (text fig. 2-A), there are three to four (except area 5 which has two) post-basicoronal plates in each interambulacral column, and one large and three to four smaller plates in each ambulacral column on the oral surface. *D. gibbsii* (text fig. 2-C) is similar except that there may be an extra plate in each column of the anterior areas and only two in each column of interambulacrum 5—a feature caused by the greater excentricity of this species. The new species, *D. elsmerensis*, is generally similar to *D. excentricus*.

In *Merriamaster*, the arrangement of the plates is quite similar to *Dendraster* except that the basicoronal plates are proportionately a little larger, and that there are more plates in the columns on the oral side and the individual plates near the ambitus have a very low altitude in proportion to their length, a character in which they are allied to *Astrodapsis*. In the ambulacral areas, there is one large post-basicoronal plate followed by four to six plates in a column, with the most plates in the anterior areas, while there are three to five post-basicoronal plates in the columns of the interambulacral areas, again with the most plates in the anterior areas.

The internal partitions and buttresses of the four genera are shown on plate 1, figs. 5, 7, 8 and on plate 2, figs. 1-4. All the genera are typified by two fairly closely spaced radial partitions extending from the periphery inwards some distance in each interambulacrum, together with a variable development of supplementary concentric buttresses. Those of *Merriamaster* (pl. 2, fig. 3) are the simplest, extending in only about one-third of the distance from the periphery and having only a slight development of supplementary buttresses, this latter fact separates it from *Astrodapsis*, which in the species examined (no members of *A. whitneyi* stock) has no supplementary buttresses. *Anorthoscutum* (pl. 1, figs. 5, 7) has the radial partitions and buttresses about equivalent to *Merriamaster* but much heavier. *Echinarachnius* (pl. 1, fig. 8) has the same radial partitions, and finer but more abundant buttresses covering a greater area of the floor, but not as many as in all of the Dendrasters examined except *D. elsmerensis*. Typical *Dendraster* (pl. 2, figs. 1, 4) has exceedingly abundant buttresses but *D. gibbsii* has fewer but exceedingly heavy buttresses covering nearly as much area as those of the typical species.

For comparison with the preceding genera, specimens of *Astrodapsis arnoldi* var., *Echinarachnius gabbii* (Rémond), *E. merriami* (Anderson), *E. gabbii kleinpelli* Grant and Hertlein (*Scutella gabbii* var. *tenuis* Kew), *E. vaquerosensis* (Kew), and *Eoscutella coosensis* (Kew) were examined.

Astrodapsis is readily separable from the genera previously discussed in this paper by the numerous low plates of the ambulacral columns outside of the petals extending over the ambitus onto the oral surface before they begin to increase in size, in addition to having raised petals, and ambulacral grooves on the oral surface of the general type of *Echinarachnius parma*. In the arrangement of the plates, the internal partitions, and the slightly raised petals, *Merriamaster* appears to be much closer to *Astrodapsis* than to *Dendraster* where *Stewart* (Woodring, Stewart and Richards, 1940, p. 81) placed it.

Echinarachnius vaquerosensis has rather large post-basicoronal plates in contact with the basicoronal interambulacral plates only in areas 2 and 3, in much the same manner as *E. parma*. The pattern of the ambulacral grooves cannot be made out, but the species, except for its laterally elongated shape, appears to be well placed in this genus. In

E. gabbii, the post-basicoronal interambulacral plates are in contact with the basicoronal row in each area except 5, but the basicoronal plates are rather large, and several ambulacral plates of very low altitude in comparison to their width are present on the oral surface. The first post-basicoronal interambulacral plates are somewhat elongated, and it is possible that it may be near the ancestral stock of *Anorthoscutum*, but certainly this species is not in the direct line for it has too many plates in the ambulacral columns outside of the petals. It may also be a side branch of the stock that produced *Astrodapsis*. Certainly it is not a typical *Echinarachnius*, although Nisiyama (1940, p. 817) included it within the typical subgenus.

In *Echinarachnius gabbii kleinpelli* (hypotype no. 11370), the basicoronal plates are moderately large and the post-basicoronal plates in interambulacral areas 2 and 3 are closer (just slightly separated) to the basicoronal plates than those of the other areas, indicating that it is apparently the start of a side branch of the main *Echinarachnius* stock. It is possible that it could be the ancestral radicle of *Dendraster*, which by further separation of the interambulacral plates in areas 2 and 3, slight reduction in size of the basicoronal plates, coupled with a posterior movement of the center of the apical system, evolved into the typical *Dendraster*. Unfortunately the pattern of the ambulacral grooves in this species is unknown. It cannot be closely related to *E. gabbii* because it lacks the numerous small ambulacral plates in the areas outside the petals and therefore should be known as *E. kleinpelli*, not *E. gabbii kleinpelli.*

Echinarachnius merriami (hypotype no. 32984) has rather large plates in the basicoronal row, and the tip of the primordial basicoronal interambulacral plates just touching the tip of subsequent plates in each area except 5. Unfortunately the available specimens of *E. merriami* are not adequate for complete description of the plate systems, but it appears possible that it could be ancestral to all later scutellids discussed in this paper. Nisiyama (1940, p. 816) places this species in his subgenus *Kewia* which he considers the ancestral radicle of the Echinarachnid echinoids.

With regard to the anomalous position of the periproct in *Dendraster elsmerensis*, and that noted by Stewart (Woodring, Stewart and Richards, 1940, pp. 81, 83) for the "small variety" specimens of *D. coalingensis*, it is not nearly as anomalous as might be supposed. The very young of all Scutellids have the periproct either marginal or on the aboral surface. Young *Echinarachnius parma* continue to have the periproct on the aboral surface until a diameter of about 15 mm. (Agassiz, 1870, pt. II, pl. 12, fig. 13). At a diameter of 10 mm., young *D. excentricus* have the periproct just slightly submarginal, and at a diameter of 4 mm., it is marginal. It may be assumed, therefore, that *D. elsmerensis* is primitive in the position of the periproct, while the occasional specimens of *D. coalingensis* mentioned by Stewart are merely individuals in which the migration of the periproct to the oral surface has been retarded.

In summary of the above discussion, it is clearly apparent that *Anorthoscutum* is not as closely related to *Echinarachnius* as was suggested by Stewart (Woodring, Stewart and Richards, 1940, p. 83), and that both genera are distinct from *Dendraster*, although an upper Miocene species of *Echinarachnius*, *E. kleinpelli*, could have given rise to *Denraster*. *Merriamaster* probably is not as closely related to *Dendraster* as it is to *Astrodapsis*.

The genus *Dendraster* may be considered as characterized by an excentric apical system, small plates in the basicoronal row, post-basicoronal interambulacral plates not in contact with basicoronal plates, periproct usually on oral surface, and unequal bifurcating ambulacral furrows extending onto aboral surface in posterior areas. *Echinarachnius* is characterized by a normally central apical system, very large plates in basicoronal row, in advanced species with interambulacral plates in areas 2 and 3 just barely in contact with basicoronal row, but in primitive species, areas 1 to 4 may be in contact with basicoronal row, ambulacral furrows simple, trifurcating near margin and all equal, periproct usually marginal. In *Anorthoscutum*, as far as known, the periproct is always supramarginal, the apical system is usually somewhat excentric, the ambulacral furrows approach those of *Dendraster* in complexity, but the major branching is a modified trifurcation instead of bifurcation. *Merriamaster* is intermediate between *Dendraster* and *Astrodapsis*, having small plates in the basicoronal row, the post-basicoronal interambulacral plates separated from the basicoronal plates, a rounded margin, submarginal periproct, a few internal buttresses, and simple branching ambulacral furrows which may extend onto the aboral surface.

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MUSEUM OF PALEONTOLOGY
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DISCUSSION

NATIONAL RESEARCH COUNCIL COMMITTEE ON THE INTERRELATIONS OF PLEISTOCENE RESEARCH

For some time, research workers in such fields as prehistoric archeology, climatology, oceanography, ecology, pedology, as well as certain branches of geology, have been aware of their need for a knowledge of recent progress in each of the other fields so closely allied to their own. The common ground for all is a broad knowledge of Pleistocene history. But as each of the sciences has its own mode of approach, its contributions have remained largely unknown outside its own field. To provide the much-needed coordination and intercommunication, the National Research Council established in June, 1947, a Committee on the Interrelations of Pleistocene Research under the chairmanship of Professor Richard Foster Flint of Yale University. The other members of the committee are: Edward S. Deevy, Jr. (biogeography) Department of Biology, Yale University; Loren D. Eisley (anthropology and new-world archeology), Department of Anthropology, University of Pennsylvania; Claude W. Hibbard (vertebrate paleontology), Museum of Paleontology, University of Michigan; Chauncey D. Holmes (glacial erosion and sedimentation), Department of Geology, University of Missouri; Helmut E. Landsberg (meteorology and climatology), Committee on Geophysical Sciences, Research and Development Board, Washington, D. C.; Hallam L. Movius, Jr. (old-world archeology), Peabody Museum of Archeology and Ethnology, Harvard University; Fred B. Phleger, Jr. (oceanography and sea-floor geology), Department of Geology, Amherst College; Louis L. Ray (glacial stratigraphy; alpine glacial geology), U. S. Geological Survey, Washington, D. C.; H. T. U. Smith (eolian features; frozen ground; stream terraces), Department of Geology, University of Kansas; and James Thorp (soil science), Division of Soil Survey, U. S. Department of Agriculture, Lincoln, Nebraska.

Individual preliminary work was begun as soon as the committee was authorized, and on April 5 and 6 the committee held its first meeting. The results are most gratifying. In following out the program planned at that time, the committee invites suggestions and assistance from all scientists interested in problems of the Pleistocene. Correspondence should be directed to the individual member representing the particular field, or to the chairman.

CHAUNCEY D. HOLMES

DEPARTMENT OF GEOLOGY
UNIVERSITY OF MISSOURI
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REVIEWS

Atomic Energy, being the Norman Wait Harris Lectures delivered at Northwestern University by KARL K. DARROW, Ph.D. Bell Telephone Laboratories. Pp. 80. New York, John Wiley and Sons, London, Chapman and Hall, 1948, $2.00.—The requirement that a series of lectures be published sometimes ruins them for their audiences, as it is an excuse for the speaker to read a book supporting or advancing his reputation for originality. In the present case the lecturer was too honest to make hay in this manner while his hearers slept. It is even probable that he was as much enjoyed by those present as he will be by the "layman" to whom he hesitantly addresses his preface.

The physicist, who is not particularly invited to read this little book, will find nothing in it to offend his critical eye except the statement that 1913 was the year in which the existence of isotopes with different masses was established. It is literally correct, but the idea of isotopes came six years earlier with the announcement by Boltwood of the discovery of ionium. A chemist may, and should, object to the statement that hydrogen is explosive. (Even its "demassing" into helium is a slow process.) All experimentalists will be slightly shocked, as was this reviewer, by the author's relative neglect of the techniques in nuclear physics. A wave of the hand to a Van de Graaf machine, a few nods to C.T.R. Wilson's cloud chamber, and an oblique glance at scintillation counting are about all that can be found before the moderation of a nuclear reactor is described. The zealot for one world or none, who frequently does not make it too clear which horn of this dilemma he prefers to be impaled upon, will not like this book because it does not spend many words on *The Bomb,* and perhaps, relies too much upon the dubious emphasis of understatement (p. 61). The intelligent man who can appreciate quantitative reasoning will find this a pleasant way to find out *where* the fantastic energies of nuclear reactions arise, and if he doesn't quite see *how* mass is turned into kinetic energy he can comfort himself by learning here that nobody is yet appreciably wiser on this point. L. W. MCKEEHAN

Electrical Measurements in Theory and Application; by ARTHUR WHITMORE SMITH. Fourth Ed. New York, Toronto, London, 1948 (McGraw Hill Book Co., $4.25).—The first edition, entitled Principles of Electrical Measurements, appeared in 1914, and already was backed by ten years of personal experience. Throughout the series the author has included relatively lengthy discussions of the units in which results of measurement are obtained. The theory of measurement, in any more fundamental sense, is not considered, and the inclusion of the word "theory" in the title is therefore somewhat misleading. It is also confusing that the author elects to begin building up the practical system of units with the law of force between current elements, so that, as compared with more usual procedures, the definitions given as basic are strikingly arbitrary in appearance. Perhaps the most disturbing change since the third edition, of 1934, is the adoption of the half-baked notion, introduced into quick courses for electronics technicians during World War II, that because electrons almost visibly flow in some circuit elements, the positive direction of electric current must be inverted from its classical sense to follow the motion of negative electricity. This is not to be accepted on any less authority than that of an international convention and is especially unattractive at a time when the motions of positively charged particles are being so extensively studied.

The descriptions of experiments are painstakingly complete, and the book will be found a satisfactory laboratory manual where direct current and low-frequency measurements are to be emphasized. The amount of space devoted to high frequency electrical measurements and to modern electronic devices, is trivial. (The cathode-ray oscilloscope, and the electron-tube amplifier, are brushed off in a total of two pages, and electrical counting is not even mentioned.)

The circuit diagrams (except for arrows pointing up potential gradients) are good. Other illustrations, especially half-tones from advertising brochures, are less useful. Paper and printing are excellent. L. W. MCKEEHAN

Very High Frequency Techniques, Volumes I and II; by THE STAFF OF THE RADIO RESEARCH LABORATORY, HARVARD UNIVERSITY. Pp. 1059; 923 Figs. New York, 1947 (McGraw Hill Book Co., $14.00).—The two volumes contain over one thousand pages, with nearly one thousand illustrations, divided into thirty-five chapters. The table of contents indicate that thirty to forty authors had some part in the production of the publication, and most of these people have already demonstrated their ability in previous publications. Under the able editorial direction of Dr. Herbert J. Reich they have produced an amazingly concise, yet complete, reference on the work done by the staff of the Radio Research Laboratory during the war years. The illustrations are, for the greater part, line drawings that serve to illustrate important features better than photographs, in that interiors can be indicated.

Antennas and microwave tubes are emphasized in these volumes with special emphasis on the development of the resnatron and on the characteristics of certain types of magnetrons. An outline of the method of measurement is given in nearly every instance of supplied data, enabling the reader to qualify the data to his own satisfaction. In addition to the large amount of data on characteristic types, theoretical information is also supplied. The section on impedance matching and broadbanding (Chapter 3) is typical of the clear presentation of rather difficult concepts encountered throughout both books.

A bibliography is supplied in supplementary form. It is quite complete, but of necessity bears upon technical reports that may not be readily available.

These books should be of value as a text in advanced high-frequency classes and as references for research and development workers already established in the field. WILLIAM L. BRISCOE

Heat Conduction, with Engineering and Geological Applications; by LEONARD R. INGERSOLL, OTTO J. ZOBEL and ALFRED C. INGERSOLL. Pp. viii, 278; 43 figs. New York, 1948 (McGraw-Hill Book Co., Inc., $4.00).—This is a thorough revision and amplification of the text by Ingersoll and Zobel first published in 1913. The basic mathematical theory of heat conduction, dating back to Fourier, is presented in some detail. For those whose interests are largely on the practical side there is a valuable chapter on auxiliary graphical and other approximation methods for solving heat conduction problems with only the simplest mathematics. These procedures are often of sufficient accuracy for practical purposes, and provide solutions for many problems too complex to be treated by rigorous classical methods. Only a reasonable knowledge of calculus is necessary for reading this book.

Geologists, geographers and engineers will find discussions of many applications of heat conduction in their fields. References to the literature are numerous and up-to-date. This is definitely an authoritative book on heat conduction, of value both to students of mathematical physics and to all scientists who need these methods in the solution of their practical problems. W. W. WATSON

General Chemistry; by FRANCIS E. RAY. Pp. viii, 592; 183 figs. Chicago, 1947 (J. B. Lippincott Co., $4.50).—At the occasional risk of the contempt of students for the too familiar, this book brings chemistry very much to life for the beginner. Elementary enough in its approach for nearly any freshman course, it nevertheless contains enough new material to hold the interest even of students for whom most of the material will be in the nature of review.

The language is refreshingly simple, with none of the much-too-common flights of pedagogical verbosity. There is an inquisitive attitude in the introductions to principles which is bound to stimulate interest. The author's practice of sketchily covering historical matter and carrying this up to hints of present boundaries of knowledge will have a similar effect.

In describing processes of chemical manufacture, the underlying principles are usually mentioned in the opening sentence rather than vaguely referred to at haphazard points in the description. Further, up-to-date processes are described with an ample supply of recent photographs and diagrams. ELLIOT PIERCE

Organic Chemistry; by FRANCIS E. RAY. Pp. x, 750. Chicago, 1947 (Lippincott Co., $4.50).—The author's stated purpose is to bridge the gap between text book and current literature. Though his efforts in this direction are heroic, it is as difficult a task for a one-volume elementary text as for a one-year elementary course. Actually, while falling short of this goal, he has produced an excellent text for an organic course for pre-medical students or for an introductory course for chemists.

The book is divided up according to classes of compounds, broken down as far as possible. A summary of synthetic methods appears near the end of each chapter. A unique feature is the use early in the book of photographs of molecular models. This should encourage early comprehension of three dimensional molecules. It should also suggest to instructors the value of using molecular models in class. Properties of compounds are tabulated more frequently and more plainly than in most texts, another useful feature.

ELLIOT PIERCE

Frontiers in Chemistry. Vol. V. Chemical Architecture; by R. E. BURK and OLIVER GRUMMITT. New York, 1948 (Interscience Publishers, Inc., $4.50).—The nature of *Frontiers in Chemistry* has been discussed before in these columns. Volume V of the series is a particularly important and interesting contribution, containing as it does a collection of highly authoritative discussions of molecular structure. The authors of the articles are, as usual, well known workers in their fields. The contents are as follows:

Hugh S. Taylor: *Applications of Molecular Geometry in the Field of Reaction Mechanism.*
Charles P. Smyth: *Dipole Moment, Resonance and Molecular Structure.*
W. Conrad Fernelius: *Structure of Coordination Compounds.*
B. E. Warren: *X-Ray Studies of Randomness in Various Materials.*
H. Mark: *Light Scattering in Polymer Solutions.*
Miroslav W. Tamele: *The Nature of Inorganic Gels.*

The discussions themselves, particularly in view of the excellent bibliographies appended to each, might well serve as a text for a large section of a graduate course in general chemistry. Any chemist will find this volume interesting, informative, and useful.

HENRY C. THOMAS

A Textbook of Geomorphology; by PHILIP G. WORCESTER. Pp. vii, 584; 385 figs. New York, 1948 (D. Van Nostrand Company, Inc., $5.25).—This second edition of Worcester's textbook is little changed from his first edition. The most notable addition, which will be of considerable help in teaching, is the lists of topographic maps at the end of eight of his nine chapters dealing with topography of land forms. Recent publications added bring the reference lists at the end of the chapters up to date. The chapter "Materials of the Lithosphere. Elements and Minerals." in the first edition now has been reduced to an appendix. This appropriately might have been done to the chapter "Materials of the Lithosphere. Rocks." The section on Characteristics of Streams, which includes descriptions of the types of streams, such as consequent, subsequent, obsequent, insequent, antecedent, and superimposed streams (pp. 146-149), is still the best of its kind found in Geomorphology textbooks. In the chapter "Shore Forms and Shore Processes of Seas and Lakes," a new section on Rip Currents (p. 381) has appeared in place of a paragraph on Rip Tides and Rip Currents, with a new diagram (fig. 265) to illustrate. There are new photographs, two of which are oblique aerial photographs (figs. 339 and 341) of a coral atoll and a volcanic island with fringing and barrier reefs. More photographs of this nature would be welcome. Some of the pictures have lost their original charm and detail through repeated use. This book still serves as a good reference and fulfills its main purpose of providing a well organized textbook in elementary geomorphology.

SIDNEY E. WHITE

A Hadrosaurian Dinosaur from New Jersey; by EDWIN H. COLBERT. Proceedings of the Academy of Natural Sciences of Philadelphia, Vol. C. 1948, pp. 23-37. In this brief paper Doctor Colbert announces the discovery of a Cretaceous dinosaur which he refers to the species *Hadrosaurus minor* described by Professor Marsh in 1870. Our knowledge of the New Jersey hadrosaurs is very meagre for it is based upon several specimens found from time to time in the marl pits as a byproduct of commercial activity, none of which approaches the degree of completeness of the numerous hadrosaurs found in the west. As a consequence any material which may be studied by so competent a paleontologist as Doctor Colbert has an actual significance far beyond its apparent value as a specimen. The present partial specimen, consisting as it does of thirteen vertebrae, a pubis, ischium, femur, and part of a fibula, adds certain diagnostic characters to our understanding of the animal. Unfortunately, however, no part of the skull was found, which, were it present, would greatly amplify our knowledge of these interesting forms.

The specimen was recovered from a sand pit near Sewell, near Barnsboro, New Jersey, the geological horizon being the Navesink formation, Cretaceous. It is now in the Museum of the Academy.

RICHARD S. LULL

Farm Soils, Their Management and Fertilization; by EDMUND L. WORTHEN. 4th Ed. Pp. xiii, 510; 199 Figures. New York, 1948 (John Wiley and Sons, Inc., $3.20).—The present edition was published seven years after the third edition. It is a book prepared for farmers and others interested primarily with practical soil management problems. The chapter titles are as follows: Selecting the Soil and Planning Its Management, Growing the Crop, Draining and Irrigating Soils, Controlling Soil Erosion, Tilling the Soil, Manuring, Liming, Fertilizing, Replenishing Nitrogen and Supplying Organic Matter, Managing Field Soils, Managing Pasture Soils, Managing Garden, Greenhouse, and Lawn Soils, and Managing Fruit Soils. The chapter on controlling soil erosion is a new one and much useful information is given on this subject. The chapter on Replenishing Nitrogen and Supplying Organic Matter has been revised from the previous edition and new material has been added.

One criticism of the book is that only meager information is given about the soil itself. The book is woefully lacking in information about kinds of soils occurring in the United States and about their development. The information that is given is archaic, and anyone interested in what soils are would have to refer to another book to find out. Knowledge of this kind is important for county agents, agricultural extension workers, and soil conservation technicians use it in their every-day recommendations on soil management.

Little information is contained on the newer kinds of fertilizers produced and especially the newer nitrogen forms which were in use a year or so prior to the publication of this book. Little information is contained on soil structure and ways for managing or improving the physical condition of the soils.

For one who is interested in practical farming this book should prove very useful. Most of the material as well as the pictures has been brought up to date. It is easy to read and understandable. It is recommended to those interested in making a living off of a farm or those interested in managing farms. C. L. W. SWANSON

Soil Physics; by L. D. BAVER, 2nd Ed. Pp. xiii, 398; 89 figures. New York, 1948 (John Wiley and Sons, Inc., $4.75).—This expanded and revised edition of Baver's book on soil physics contains 28 more pages than the first edition. The chapter titles remain the same as in the first edition. They are: Introduction, Mechanical Composition of Soils, Physical Characteristics of Soil Colloids, Soil Consistency, Soil Structure, Soil Water, Soil Air, Soil Temperature, Physical Properties of Soils and Tillage, and Physical Properties of Soils in Relation to Runoff and Erosion.

New material has been included on electron micrographs and clay particles, effect of micro-organisms on soil structure, plowing, soil water, and raindrop effects on soil erosion. Addition of this new material makes this book more than ever the leading text book on soil physics.

For those interested in soil moisture relationships this subject is discussed in detail. Soil moisture energy relationship are adequately discussed including the pF concept of soil water.

This book brings together and discusses the relationship existing between structure as it exists in the natural soil profile and structure as we know it in cultivated agricultural soils. Too often the tendency is to treat these two kinds of soils structure separately not relating one to the other.

It is unfortunate that the author did not go into the mineralogical nature of clay in greater detail even though he says it is not the purpose of the book to discuss this matter thoroughly. The reviewer believes that, had he drawn upon the large store of mineralogical information on clays available in the mineralogical literature, his treatment of clays and their relationship to soils would have been improved. Even so, his discussion of soil clays brings together in one place more information about them than is available elsewhere.

For anyone interested in soil physics this is a must book. It is well written and includes a good list of references on the field of soil physics. C. L. W. SWANSON

Human Geography, An Ecological Study of Society; by C. L. WHITE and G. T. RENNER. Pp. x, 692; 437 figs. New York, 1948 (Appleton-Century-Crofts Inc., $6.00).—The content of the introductory course in geography is a topic for endless debate. Some hold for rigid earth-science, with little or no human geography. Others urge economic geography, perhaps with a brief earth-science introduction. A third group recommends world regions as the first geographical study. White and Renner follow a fourth tradition, that of "human ecology."

Because each chapter deals first with a natural factor and then with human "adjustments," the book will be poison to those who eschew even the most dilute environmentalism. Geography, it has been truly said, cannot be defined as a field of knowledge in terms of a particular kind of explanation. But an introductory textbook is not a methodological treatise. The ecological approach, if handled with skill and with candor as to its limitations, may give the beginning student an illuminating introduction to geography which is not necessarily wrong because it is incomplete.

The reviewer would criticize White and Renner for their deviations from the ecological approach rather than their adherence to it. The chapters are interesting and convincing in their descriptions of the habitats and the simpler human adjustments to them. The limitations of the ecological approach become apparent when the complexities of society and politics are reached. Some of the "political adjustments" are not adjustments at all, but merely coincidences. Others are highly debatable, such as the statement that irrigation in the Middle East required absolute despotism (p. 419).

Every man has his home in a climatic region, surrounded by landforms, on some kind of soil, and near surface or underground water. Relatively few men live in, on, or even near important economic mineral deposits. For most men, economic minerals become part of the effective environment only when they are extracted and, in most cases, processed. They are cultural rather than natural factors. For this reason, the strict ecological approach breaks down when minerals are studied and these chapters of White and Renner seem lifted from economic geography. There is no doubt, however, that the book would have been criticized had minerals been ignored or treated only briefly.

Specialists will find a number of errors. We are told, for instance, that coal occurs in veins or sheets (p. 475) and that "Nature intended that the Mediterranean should be an international sea" (p. 201). The controversial Wegener theory of continental drift, Taylor theory of race migration, and organic theory of society are espoused with little or no discussion of alternative views. A conscientious teacher will have to point out overgeneralizations. Nevertheless, the book is readable, as textbooks go, and it may be better for students to read a book that falls short of complete accuracy rather than merely possess one that is accurate but hopelessly dull. Excellent illustrations, clear maps, well-spaced lines, and well-designed pages are additional attractions.

STEPHEN B. JONES

PUBLICATIONS RECENTLY RECEIVED

Silva of North America, a description of the trees which grow naturally in North America exclusive of Mexico; by Charles Sprague Sargent. Reprinted, 14 volumes (bound in 7). New York, 1948 (Peter Smith, $200.00).

Man in the Modern World; Julian Huxley. New York, 1948 (Mentor Book, New American Library of World Literature, $.35).

Atomic Energy; by K. K. Darrow. New York, 1948 (John Wiley and Sons, $2.00).

Climatic Accidents in Landscape Making; by C. A. Cotton. Second printing. New York, 1948 (John Wiley and Sons, $7.00).

Hawaii Division of Hydrography. 1947. Bulletin 12. Geology and Ground-Water Resources of the Island of Niihau, Hawaii; by H. T. Stearns. Petrography of Niihau; by G. A. Macdonald.

Virginia Geological Survey. Bulletin 63. Geology and Ground-Water Resources of the Coastal Plain in Southeastern Virginia; by D. J. Cederstorm. University, 1945.

Electrical Measurements in Theory and Application; by A. W. Smith. New York, 1948 (McGraw-Hill Book Co., $4.25).

Bibliography of the Literature on the Minor Elements and Their Relation to Plant and Animal Nutrition. Fourth Edition. Vol. 1. New York, 1948 (Chilean Nitrate Educational Bureau, Inc.).

Bibliography of the Literature on Sodium and Iodine in Relation to Plant and Animal Nutrition. First Edition. Vol. 1, New York, 1948 (Chilean Nitrate Educational Bureau Inc.).

Illinois Geological Survey. Report of Investigations No. 134. Differential Thermal Curves of Prepared Mixtures of Clay Minerals; by R. E. Grim. Rehydration and Dehydration of the Clay Minerals; by R. E. Grim and W. F. Bradley. Bulletin No. 57. Developments in Illinois and Indiana in 1947; by A. H. Bell and R. E. Esarey. Urbana, 1948.

Chemistry of the Carbohydrates; by W. W. Pigman and R. M. Goepp, Jr. New York, 1948 (Academic Press, Inc., $10.80).

Readings in the Physical Sciences; edited by H. Shapley, H. Wright and S. Rapport. New York, 1948 (Appleton-Century-Crofts, Inc., $3.00).

The Solar System Analysed. Old Laws from a New Angle; by F. C. Attwood. Auckland, New Zealand, 1947 (The Dawson Printing Co., Ltd.).

Farm Soils, Their Management and Fertilization. Fourth edition; by E. L. Worthen. New York, 1948 (John Wiley & Sons).

Outlines of Physical Chemistry; by F. Daniels. New York, 1948 (John Wiley & Sons, $5.00).

Physical Geology, third edition; by C. R. Longwell, A. Knopf and R. F. Flint. New York, 1948 (John Wiley & Sons, $5.00).

The Essentials of Organic Chemistry by C. W. Porter and T. D. Steward. Boston, 1948 (Ginn & Co., $4.00).

Nuclear Radiation Physics; by R. E. Lapp and H. L. Andrews. New York, 1948 (Prentice-Hall, Inc., $4.50).

Virginia Geological Survey. Bulletin 65. Industrial Limestones and Dolomites in Virginia: Northern and Central Parts of Shenandoah Valley; by R. S. Edmundson. University, 1945.

U. S. Geological Survey: Bulletins as follows: 954-B. Geology and Chromite Deposits of the Camagüey District, Camagüey Province, Cuba; by D. E. Flint, J. Francisco de Albear and P. W. Guild. 954-C. Aluminous Lateritic Soil of the Republic of Haiti, W. I.; by S. S. Goldich and H. R. Bergquist. Price $.40. 859-A. Geophysical Abstracts 132 January-March 1948; by V. L. Skitsky and S. T. Vesselowsky. Price $.25. 858-B. Geophysical Abstracts 133 April-June 1948 (Numbers 9962-10212); by V. L. Skitsky and S. T. Vesselowsky. Price $.25. Professional Papers as follows: 213 Gold Deposits of the Southern Piedmont; by J. T. Pardee and C. F. Park, Jr., 213 Maps. Washington, D. C., 1948.

American Journal of Science

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THE FLOOR OF THE SHENANDOAH VALLEY[1]

PHILIP B. KING

ABSTRACT. The floor of the Shenandoah Valley near Elkton, Va., is commonly supposed to be the dissected surface of the Valley Floor or Harrisburg peneplain. Actually, the principal element of the floor is a series of gravel-capped benches, formed in Pleistocene time during alternating periods of erosion and deposition, and rising step-like toward the bordering mountains away from the flood plain of the South Fork of the Shenandoah River. Unconformably beneath the gravels is residual clay, resulting from the decay of the bedrock of Cambrian and Ordovician limestone, dolomite, and shale. The clay is thickest on the southeast side of the valley, where it attains 200 feet. It probably formed during a single epoch in the Tertiary, under mild and humid climatic conditions, toward the close of a period of peneplanation—that is, near the time of completion of the Valley Floor peneplain. At one locality a small body of ancient gravel, probably of Tertiary age, lies stratigraphically between the residual clay and the Pleistocene gravels, and is unconformable with each. The Valley Floor peneplain is believed by some authors to have been formed in late Tertiary time, but the sequence of events implied by the geological materials of the Elkton area suggests that it is considerably older.

INTRODUCTION

A GREAT deal that has been written on the geomorphology of the Appalachian area has to do with land forms and surfaces, and much less has been written on the geological materials of which these surfaces are composed. This paper deals with the geological materials which compose the land forms of one area in the Appalachians, that near the town of Elkton in the Shenandoah Valley of northern Virginia. Although many details are uncertain, it is the writer's belief that the Cenozoic history was more complex than has generally been supposed, and probably different from what one would conclude from a study of the land forms alone.

In 1940 and 1941, the writer and John Rodgers, now of Yale University, mapped geologically a belt of country 20 miles along the northwest foothills of the Blue Ridge and the southeast edge of the Shenandoah Valley in the vicinity of Elkton, Va. The investigation was a part of the wartime strategic minerals program of the U. S. Geological Survey, and the purpose was a search for reserves of manganese ore. The investigation included study of stratigraphy and structure of the bedrock formations, of manganese-bearing residuum which overlies some of the bedrock, and of superficial deposits that overlie the residuum and bedrock. During the same period the U. S. Bureau of Mines prospected two tracts in the area by means of drill holes and shafts, and their results were studied geologically by the writer. The investigation was therefore focused on surface and near-surface geological materials, and provided an exceptional opportunity to correlate these materials with the land forms.

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[1] Published by permission of the Director, U. S. Geological Survey.

The economic aspects of the investigation in the Elkton area were set forth in a strategic minerals bulletin (King, 1943), and the general geology has been described in a manuscript that has recently been submitted to the Geological Survey for publication (King, 1947). This paper is a summary of one chapter in the latter manuscript.

GENERAL GEOGRAPHY AND GEOLOGY

The Elkton area lies in Page and Rockingham Counties, Va. The town of Elkton is 20 miles south of Luray and 18 miles east of Harrisonburg. The area is drained by the South Fork of the Shenandoah River, which follows the axis of a gently sloping valley—the southeastern branch of the Shenandoah Valley—from 4 to 7 miles wide, which stands at an altitude of 800 to 1,000 feet (map, figure 1). The valley is bordered on the southeast by the Blue Ridge and its fringe of foothills, whose summits rise to altitudes of nearly 4,000 feet. It is bordered on the northwest by the narrow hogback ridges of Massanutten Mountain, the highest of which reach an altitude of 3,300 feet.

The bedrock of the area consists of formations of pre-Cambrian, Cambrian, Ordovician, and Silurian age, the oldest lying to the southeast and the youngest to the northwest. Table 1, below, shows the general succession insofar as it bears on the discussion that follows:

TABLE 1

Generalized Section of Bedrock Units in Elkton Areas, Virginia.

Age	Formation	Thickness (feet)	Lithologic character	Topographic expression
Silurian	Sandstone and quartzite	500	Sandstone and quartzite	Ridge crests of Massanutten Mountain
Ordovician	Martinsburg shale	2,800	Shale and sandy shale	Northwest slope of Shenandoah Valley
	Limestone	200	Thin-bedded limestone	Axis of Shenandoah Valley; in places overlain by thin residuum
	Beekmantown dolomite	3,000	Thick-bedded dolomite	
Cambrian	Conococheague limestone	2,000	Thick-bedded, argillaceous and sandy limestone	
	Elbrook dolomite	3,000	Thin-bedded dolomite	
	Waynesboro formation	1,700	Red and brown shale; thin limestone beds	Southeast slope of Shenandoah Valley; generally overlain by thick residuum
	Tomstown dolomite	1,000	Dolomite, argillaceous dolomite, some shale	
	Antietam quartzite Harpers shale Weverton quartzite Loudoun formation	2,900	White, thick-bedded quartzite above, underlain by quartzite, siltstone, and shale	Foothills of Blue Ridge; hogbacks, dip slopes, strike ridges; much talus
Pre-Cambrian	Catoctin greenstone and underlying sediments Injection complex		Massive volcanic and plutonic rocks	Slopes and summits of Blue Ridge

The Blue Ridge is formed of massive pre-Cambrian volcanic and plutonic rocks, and its summits, standing at altitudes of 3,000 feet or more, are nearly flat or gently rolling in many places. These flat areas have been interpreted as peneplain remnants (Stose, 1919, pp. 37-38; and other authors) and have been correlated partly or wholly with the Schooley pene-

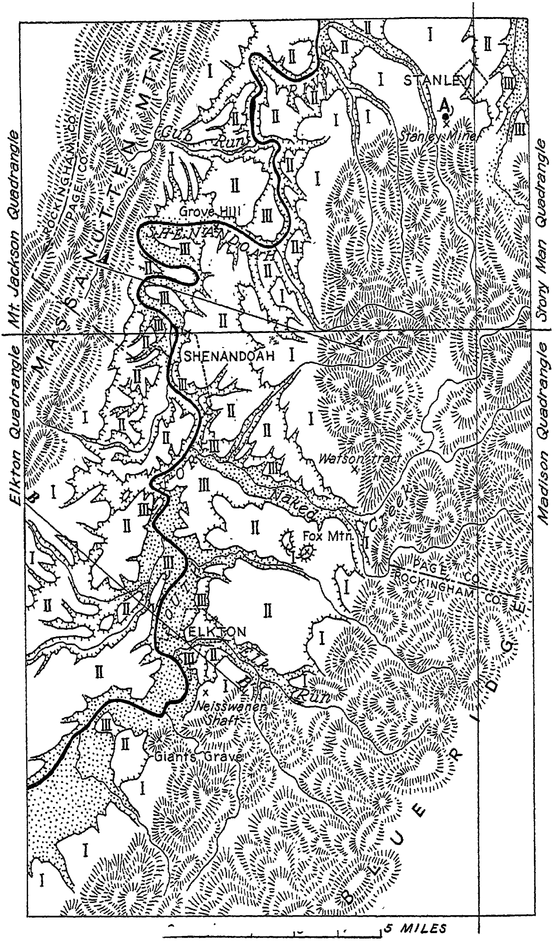
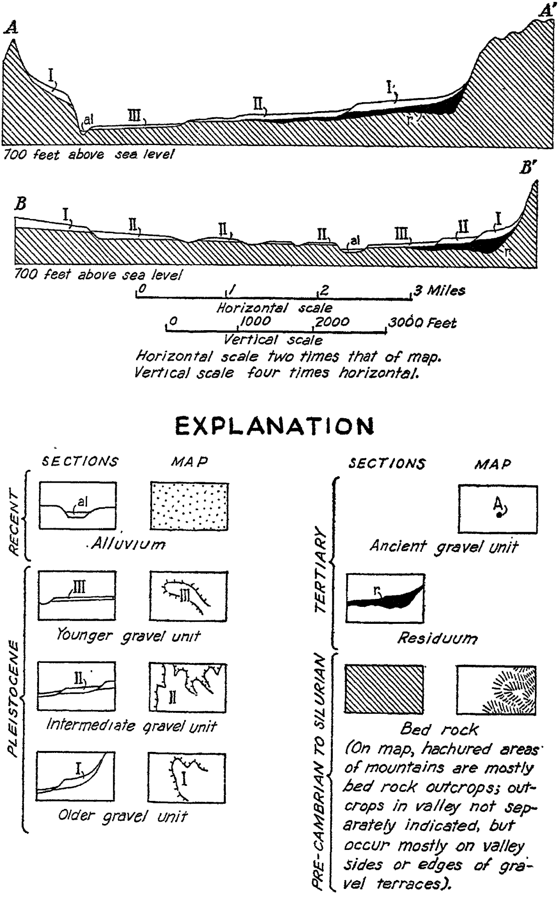


Figure 1. Map and sections of Elkton area, Virginia,



showing Cenozoic deposits, and localities mentioned in text.

plain. The crests of Massanutten Mountain have likewise been interpreted as controlled by a former high-level peneplain (Keith, 1894, p. 391; Spencer, 1897), perhaps the Schooley. Remnants of ancient erosion surfaces may well exist on the Blue Ridge, but it is uncertain how many are represented, as the flat summits stand at divergent heights in different places. If a surface once existed in the Massanutten Mountain area, it is now poorly preserved, for the mountain consists of ridges with knife-edged tops, without any flat summit areas. Below the Schooley level in parts of northern Virginia another surface, the Intermediate or Weverton peneplain, has been reported (Keith, 1894, p. 238; Stose, 1919, pp. 38-39). No evidence for this was seen in the Elkton area; the slopes intervening between the summit of the Blue Ridge and the floor of the Shenandoah Valley are steep and irregular, and there is no systematic arrangement of the foothill crests.

The area here called the "floor of the Shenandoah Valley" derives its name from the terms "valley floor" and "valley floor peneplain" which were used in older reports. Actually, it appears level only by contrast with the mountains which rise on either side. In detail it is a complex of rock-cut surfaces, of residuum, and of gravel deposits, standing at various heights above the flood plains of the present streams, and sloping gently from the bases of the Blue Ridge and Massanutten Mountain toward the axial stream in the center—the South Fork of the Shenandoah River (sections, fig. 1). Relief on the floor exceeds 500 feet in places. The surface and near-surface materials of the valley floor are indicated in table 2, below, and are described in more detail later in this paper. Rock outcrops occupy relatively small areas, chiefly in the stream valleys. By far the greater part of the surface is formed by the older, intermediate, and younger gravel units (table 2). The principal surface on the floor of the Shenandoah Valley has been interpreted as the Valley Floor, Harrisburg, or Shenandoah peneplain (Keith, 1894, pp. 374-376; Stose, 1919, pp. 39-40; Wright, 1934, pp. 28-31), a concept that will be examined at the conclusion of this paper.

RESIDUUM

Character.—The residuum of the area results from the decay of the rocks of the valley floor, and consists of insoluble

TABLE 2

Surface and Near-Surface Materials on Floor of Shenandoah Valley, Elkton Area.

Age		Name of unit	Description	Topographic relations
Quaternary	Recent	Alluvium	Gravel, sand, and clay.	Flood plains of South Fork of Shenandoah River and major tributaries.
	Pleistocene	Younger gravel unit	Thin sheet of gravel; same composition as older gravel unit.	Terraces near river, and 50-75 feet above it.
		Intermediate gravel unit	Thin sheet of gravel; same composition as older gravel unit.	Broad plains and benches on intermediate slopes of valley floor.
		Older gravel unit	Gravels, composed of pebbles and boulders, mainly of quartzite; gravel sheet thickens into coalesced alluvial fans at bases of mountains.	Narrow benches at bases of mountains, 300 to 700 feet above river.
Tertiary		Ancient gravel unit	Sand and gravel, in eroded basins in residuum, overlain unconformably by older gravel unit; mineralized.	Small areas only.
		Residuum	Clay, waxy or silty; chaotic structure; mineralized by iron and manganese oxides.	Mainly on Tomstown and Waynesboro formations, beneath older gravel unit and near foot of Blue Ridge; thin residuum on younger formations to northwest.
Unweathered bedrock of Cambrian and Ordovician limestones, dolomites, and shales.				

material, principally waxy or silty clay, which remains after the solution of limestone and dolomite, or after kaolinization of sericite in shale (King, 1943, pp. 21-26). Mixed with the clay are sand and silt, a residue of originally sandy or silty rocks, and chert, originally in the form of concretions in the carbonate rocks. Some of the clay also contains nodules, plates, and impregnations of iron and manganese oxides, which were introduced into the clay after its formation by processes of mineralization and replacement (Stose and others, 1919, pp. 44-45; King, 1943, pp. 32-35). These oxides form the iron and manganese ores of the area, and have been mined where concentrated in bodies of sufficient size. Much of the clay has a chaotic structure, resulting from collapse, compaction, and creep that took place during the reduction in volume of the original rock.

Thickness.—The residuum is thickest over the Tomstown dolomite and Waynesboro formation on the southeast margin of the valley, and is thin or even wanting over the younger carbonate formations along the axis of the valley (sections on fig. 1).

The volume of the residuum overlying the Tomstown and Waynesboro formations is indicated mainly by test holes and mine workings. Test holes put down by the U. S. Bureau of Mines on the Watson tract, 5 miles northeast of Elkton, showed a thickness of residuum of the Tomstown dolomite of 100 to 150 feet (section A, fig. 2). Test holes south of Giants Grave, 3 miles south-southwest of Elkton, showed a thickness of residuum of the Tomstown dolomite of 50 to 100 feet. The Neisswaner shaft, a mile south of Elkton, showed a thickness of residuum of the Waynesboro formation of nearly 200 feet (Hewett, 1916, pp. 61-67) (section B, fig. 2). The deepest residuum penetrated on the Watson tract extends down to the altitude of nearby Naked Creek, along which residuum is also exposed, and that in the Neisswaner shaft extends below the altitude of the nearby South Fork of the Shenandoah River.

The residuum of the younger carbonate formations farther out in the Shenandoah Valley is thinner. Thin bodies of residuum were noted above the unweathered Elbrook and Beekmantown dolomites in some outcrops, and somewhat thicker bodies were noted above the Conococheague limestone (fig. 3). In some places residuum is wanting, as along the east side of the South Fork of the Shenandoah River at Elkton, where unweathered Elbrook dolomite is overlain directly by gravels. The base of the residuum of the younger carbonate formations also seems to be higher with respect to the present drainage than that of the Tomstown and Waynesboro formations. Near Elkton unweathered Elbrook dolomite and Conococheague limestone are 50 feet or more above river level, whereas unweathered Waynesboro formation lies below river level in the Neisswaner shaft not far to the southeast. North of the town

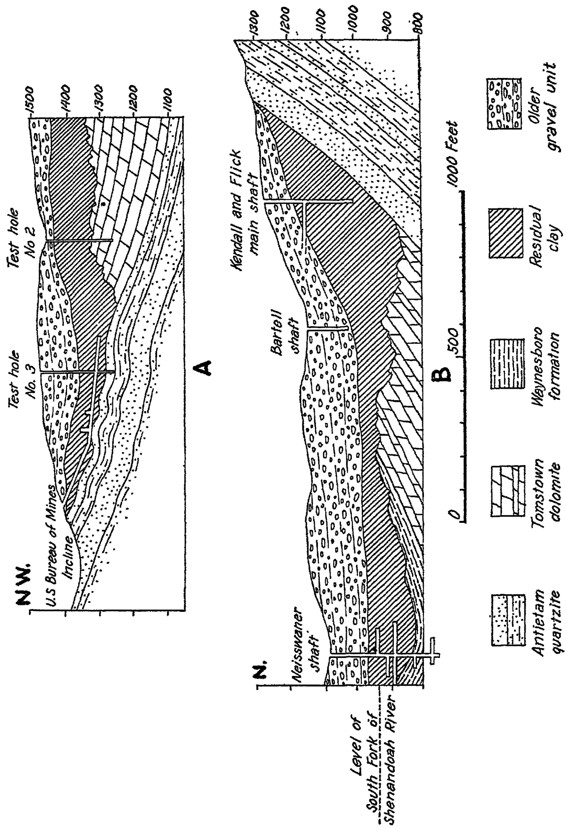


Figure 2. Sections showing inferred relations of gravel and residuum to bed rock along northwest foot of Blue Ridge. A.—On Watson tract, 5 miles northeast of Elkton. B.—Near manganese mines a mile southeast of Elkton. Section A based on Bureau of Mines drill holes and underground workings put down in 1941. Section B based on published records (Hewett, 1916).

of Shenandoah, unweathered Beekmantown dolomite extends 100 feet or more above river level.

The Martinsburg shale, northwest of the axis of the Shenandoah Valley, seems to have given rise to only small amounts of residuum, and ledges of unweathered shale extend 300 feet or more above river level.

The residuum is unconformable beneath the older, intermediate, and younger gravel units, this relation being visible in many mine workings, and in road and railroad cuts (figs. 2, 3, and 4). Open cuts of mines in residuum of the Tomstown and Waynesboro formations show coarse, cobbly or bouldery older gravels lying on the eroded surface of the residuum, and truncating the contorted structure of the residuum. The gravel contains fragments of iron and manganese oxides reworked from the residuum.

Origin.—Residuum is not being actively accumulated at the present time. Rock outcrops on valley slopes cut below the uplands of the valley floor show little or no surface accumulation of clay. Clay is no doubt being released today by weathering of carbonate rocks on the valley slopes, but under present conditions it is being carried away by erosion. The thickest accumulations of clay are beneath the upland surfaces of the valley floor, where removal by erosion would be even more active at the present time. Moreover, the clays are mineralized by iron and manganese oxides, yet many such deposits, according to Hewett (1916, p. 44)

> "are situated near the top or along the slope of isolated hills or spurs, so that since dissection of the peneplain began they have received very little of the surface drainage or ground water that might contain manganese in solution."

The unconformable relation of the residuum to the older, intermediate, and younger gravel units indicates that the collapse, compaction and creep accompanying the clay formation, and the subsequent mineralization of the clay, took place before these gravels were laid down.[2] It is therefore probable that the main accumulation of residuum took place before the deposition of the oldest gravels, and it may well have been accomplished in a single epoch.

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[2] In a previous report (King, 1943, p. 22) it was implied that accumulation of residuum took place during each erosion period which preceded each epoch of gravel deposition. This is now regarded as unlikely.

The thickest masses of residuum are those on the Tomstown dolomite and Waynesboro formation. In part this is because formations lie farthest from the axis of the valley. Their residuum would therefore be less subject to attack by later erosion than would residuum of the younger formations. However, the base of their residuum in many places extends below the level of modern drainage, and lower than the base of the residuum of formations near the axis of the valley. Very deep circulation of ground water evidently took place, for surface drainage at the time of formation of the residuum stood at a higher level than modern drainage. Along this belt of outcrop, circulating water may have been guided by the impervious beds of the Antietam quartzite which underlies the Tomstown. Similar observations and interpretations have been made by Rodgers in Bumpass Cove, Tennessee (1948, p. 15 and fig. 3).

The residuum may have accumulated in a single epoch. During this epoch, the climate may have been warmer and more humid than the present one (Hewett, 1916, pp. 46-47). Accumulation took place when the present upland areas of the valley floor lay near the grade of the streams which drained them, and when erosion was not sufficiently active to remove the residuum as it formed. In other words, conditions on the valley floor had then reached a state of peneplanation. This epoch no doubt corresponds to the time of completion of the Valley Floor or Harrisburg peneplain. The age of the Valley Floor peneplain will be considered at the conclusion of this paper.

DEPOSITS BETWEEN RESIDUUM AND OLDER GRAVEL UNIT

Lying stratigraphically between the residuum and the older gravel unit are small remnants of other deposits. Because of their position they are seldom revealed in natural exposures, but they have been uncovered in mine workings and other artificial openings. Only one remnant of such deposits was observed in the Elkton area, here termed the ancient gravel unit, but they occur widely in the southern Appalachians, as indicated below.

Ancient gravel unit at Stanley Mine.—The single observed occurrence of the ancient gravels lies in the north part of the Elkton area, in the open cut of the Stanley Mine, which is a mile southwest of Stanley at an altitude of 1,350 feet. The deposit fills a steep-sided, eroded basin in clay residual from the Tomstown dolomite, and is overlain unconformably by coarse deposits of the older gravel unit (fig. 4). The ancient deposit consists of cross-bedded red sand, in which deeply weathered, rounded pebbles of sandstone and quartzite are

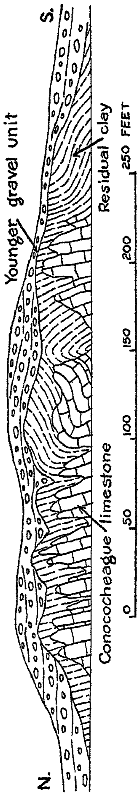

Figure 3. Section showing relation of residuum to Conococheague limestone and younger gravel unit in cut on Norfolk and Western Railway, three-fourths mile southeast of railroad station in Shenandoah.

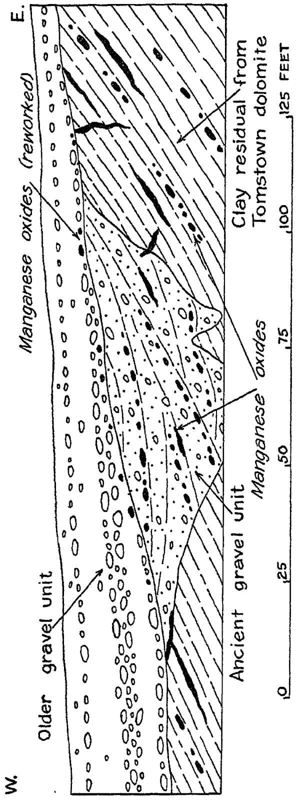

Figure 4. Section showing relation of ancient gravel unit to residuum and older gravel unit in southern open cut of Stanley Mine, 1 mile southwest of Stanley. Section is schematic, being made by combining several outcrops within the cut.

embedded. Like the underlying clay, the deposit contains nodules and masses of manganese oxides that have impregnated or replaced the original sandy sediment. The overlying older gravel unit at this locality contains pieces of manganese oxides reworked from the residuum and from the ancient gravels, indicating that at least a part of the manganese mineralization of these units took place between the time of deposition of the ancient and older gravel units.

Related deposits in other areas. — Mineralized, ancient gravels occur at the same stratigraphic position elsewhere in Virginia. Manganese-bearing pebbly sand and clay have been encountered and mined at the Kennedy Mine, Augusta County (Stose and others, 1919, pp. 103-107), and at the Midvale Mine, Rockbridge County (Hewett, 1916, p. 60), southwest of the Elkton area.

At other localities in the southern Appalachians deposits at the same stratigraphic position have been worked for bauxite. One of these occurs southwest of Greenville, Augusta County, Va., and others are found in Tennessee, Georgia, and Alabama. These were intensively studied by Geological Survey field parties under the direction of Josiah Bridge during the recent war, but most of the results have not been published. One of the deposits, near Elizabethton, Tenn., was studied by the writer and his colleagues (King and others, 1944, pp. 45, 210-213), and according to Bridge resembles those elsewhere. It lies on the valley-floor upland, and consists of a mass of kaolinitic clay with a central core of bauxite, and forms a steep-sided pocket in residuum of the underlying carbonate bedrock. Lying in the kaolin are contorted lenses of lignite. The lignite at Elizabethton, and at most other deposits, has failed to yield identifiable plant remains, but according to Bridge, plant remains collected in clay and bauxite deposits in sink holes on the valley-floor surface near Anniston, Ala., and Cedartown, Ga., have been identified by R. W. Brown as of Midway (Paleocene) age.

Age of deposits between residuum and older gravel unit.— At least two types of deposits occur between the residuum and the older gravel unit—mineralized sand and gravel, and kaolinitic bauxitic lignite-bearing clay. The origin of the deposits was diverse, and perhaps also their age, for the epoch between the time of formation of the residuum and the time of deposition of the older gravels was probably a long one. As indicated by plant remains in Georgia and Alabama, some are early Tertiary. The age of those in Virginia is unknown, but they seem most probably to have been laid down during some part of the Tertiary period.

GRAVEL DEPOSITS

Wide areas of the floor of the Shenandoah Valley are covered by gravel deposits, laid down on several surfaces that stand above the level of modern drainage. Each deposit lies on a terrace or bench, and is separated from the next by a low scarp (map and sections, fig. 1). The relations between the deposits are typically displayed southeast of Elkton, on the ridge next southwest of Elk Run, where three benches are present between the South Fork of the Shenandoah River and the adjacent mountains (section BB′, fig. 1). In some other parts of the area the differentiation is less evident; the units may merge without a distinct intervening scarp, or isolated gravel patches can be assigned only doubtfully to one unit or another. In general, however, a tripartite division of the gravels seems possible, and the units are herein referred to as the older, intermediate, and younger gravel units.

Older gravel unit.—The older gravel unit, which is the highest above the present streams, forms a series of remnant benches that fringe the northwest base of the foothills of the Blue Ridge and the southeast foot of Massanutten Mountain. In addition, two gravel-capped remnants, probably of the older gravels, stand above the plain of the intermediate gravel unit well out in the Shenandoah Valley at Fox Mountain (map, fig. 1).

The upper surface of the older gravels is probably depositional. At the outer edges of the remnants the surface stands at altitudes of 1,100 to 1,200 feet, and slopes gently toward the axis of the valley. On the mountainward sides the surface rises rapidly to heights as great as 1,500 feet and assumes the form of a piedmont alluvial slope. On the Massanutten Mountain side, where Cub Run enters the Shenandoah Valley, a fine alluvial fan is developed in the older gravels.

The deposits of the older gravel unit are probably thickest beneath the piedmont alluvial slope at the edge of the mountains. On the Watson tract, close to the mountains, test holes of the U. S. Bureau of Mines show thicknesses as great as 140 feet (section A, fig. 2). Similar thicknesses are present in the shafts of the mines a mile southeast of Elkton (Hewett, 1916, pp. 61-67) (section B, fig. 2), and in Bureau of Mines test holes south of Giants Grave, 3 miles south-southwest of Elkton. Near Stanley, in the northeast part of the area, thicknesses of 50 to 175 feet of gravel are reported by Cady (1936, pp. 80-81).

The older gravel deposit consists of pebbles, cobbles, and boulders, lying in rough layers in a sandy clay matrix. Most of the fragments are quartzite, derived on the Blue Ridge side of the Shenandoah Valley from the Antietam quartzite and other Lower Cambrian formations of the foothill belt, and on the Massanutten Mountain side from sandstone of Silurian age. Most of the quartzite fragments are rounded, indicating a certain amount of transportation, but near the apices of some of the former alluvial fans are angular blocks up to 4 feet in diameter. The older gravels do not show any evidence of mineralization, deep weathering, or other marks of extreme antiquity. Most of the component boulders are fresh and strong, although in a few places they have disintegrated to such an extent that they can be crushed into sand.

Intermediate gravel unit.—The intermediate gravel unit covers the most extensive area of the three (map, fig. 1). In places, these gravels form broad plains, as in the tract northeast of Elkton, between Elk Run and Naked Creek. As with the older gravel unit, the surface of the intermediate gravels rises toward the margins of the Shenandoah Valley and away from its axis, but alluvial fans are not developed on the mountainward sides. The surface of the intermediate gravels stands at an altitude of about 1,000 feet in the south part of the Elkton area, but descends northeastward, in harmony with the gradient of the South Fork of the Shenandoah River, to an altitude of less than 900 feet in the north part.

The intermediate gravel unit, like the older gravel unit, is composed largely of rounded quartzite fragments of various sizes. Some of the fragments may have been derived directly from erosion of the parent ledges in the mountains, others were derived no doubt from reworking of fragments in the older gravels.

A significant area of the intermediate gravels lies north of Grove Hill and northwest of the South Fork of the Shenandoah

River. Along the county road east of State Highway 12 and immediately north of the river, the intermediate gravels contain large numbers of fragments readily identifiable as having been derived from the Antietam quartzite and other Lower Cambrian formations of the foothills of the Blue Ridge. In this vicinity and throughout the whole north part of the Elkton area, the South Fork of the Shenandoah River flows in a series of entrenched meanders (map, fig. 1). The locality north of Grove Hill lies northwest of the river on the inside of one of the meander loops. The presence here of fragments derived from the Blue Ridge on the opposite side of the river to the southeast suggests that the meanders did not come into existence until after the deposition of the intermediate gravels. The meanders may have formed toward the close of the period during which the intermediate gravels were deposited.

Younger gravel unit.—The younger gravel unit lies on benches between the intermediate gravels and the present flood plain of the South Fork of the Shenandoah River (map, fig. 1). The tops of the benches are 50 to 75 feet above river level. Most of the benches are small—the largest occupying an area of about one square mile—and many lie within the loops of the entrenched meanders of the river. The whole thickness of the gravels may be observed along the edges of the benches and is less than 25 feet.. The younger gravels have nearly the same composition as that of the two preceding gravel units.

Origin and age of gravel deposits.—The gravel deposits of the Elkton area are part of an extensive system of gravels that covers much of the southeastern part of the Shenandoah Valley from the headwaters of the Shenandoah River northeastward beyond Luray (Butts, 1933, geologic map), yet little has been written on them, and the few interpretations that have been made regarding them are open to question. Stose (1919, p. 36) and Wright (1934, p. 31) mention the gravels briefly and suggest that the older were laid down during the closing stages of the formation of the Harrisburg or Valley Floor peneplain, and that the younger were river terrace deposits.

The older, intermediate, and younger gravel deposits seem to be closely related in origin. The cycle of which each succeeding deposit is a part began with dissection of the bed rock or residuum of the valley floor, and eventual reduction of part of the floor to a graded surface. During later stages of the cycle, deposits accumulated on the surface previously cut. Each cycle closed with renewed cutting by the streams to new and lower levels.

The main causes of these cycles were probably fluctuations in climate. During the initial stages of a cycle, streams were capable of transporting out of the region all the material supplied to them, because of adequate volume and because erosion was not active on the adjacent mountains. During such times, the climate was probably humid, thus permitting adequate stream flow, and protection of mountain areas from erosion by the growth of vegetation. The climate during such times was probably at least as humid as at present, and perhaps more humid.

During the later stages of a cycle, streams were incapable of removing all the material supplied to them, because of inadequate volume and because of vigorous erosion on the adjoining mountains. Butts (1940, p. 509) has interpreted the times of gravel deposition as times of greatly increased rainfall, but this view is hardly tenable. Probably the climate was dry, thus reducing the volume of stream flow and the cover of vegetation on the mountains, and exposing the mountains to attack by the forces of erosion. The climate was undoubtedly different from the present and probably was much drier, for quartzite wash is not being delivered from the mountains to the valley floor today in any such quantities as it was during the time of gravel deposition.

The successive gravel deposits are so closely related in origin that they are probably not far apart in age. Moreover, none of them exhibit marks of great antiquity, such as cementation, mineralization, or deep weathering. Very probably all the gravel deposits are of Quaternary age,[3] and were formed during the Pleistocene epoch. The fluctuations in climate herein suggested probably correspond to the fluctuations known to have taken place during the Pleistocene, the humid periods corresponding to glacial stages, and the dry periods to interglacial stages.

ALLUVIUM

Alluvium occupies the flood plains of the modern streams in the Shenandoah Valley, but it covers relatively small areas

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[3] In a previous report (King, 1943, pp. 18 and 27) the older gravels are classed as of Tertiary or Quaternary age, and the intermediate and younger gravels as of Quaternary age. There seems to be no justification for this distinction.

(map, fig. 1). From the town of Shenandoah northward the flood plain along the South Fork of the Shenandoah River is very narrow, and in places the river is enclosed by rock walls on either side. South of Shenandoah, as in the vicinity of Elkton, the alluvial areas along the river are more extensive, and in places exceed a mile in width. Here, many of the tributary streams also lie in alluvial bottoms.

The thickness of the alluvium varies. North of Shenandoah, where the South Fork of the Shenandoah River is enclosed for long distances by bedrock, riffles occur here and there in the river itself, and suggest that bedrock lies directly beneath the channel. In the broad alluvial area southwest of Elkton, bedrock is exposed in places in the channel, but at the Stonewall plant of Merck & Co., not far southeast of the river and 2 miles southwest of Elkton, water wells encounter 31, 38, 57, and 124 feet of unconsolidated deposits above the bedrock.[4] The broad alluvial flat in the vicinity of Elkton is therefore underlain by variable thicknesses of unconsolidated deposits. Some of the thicker masses may represent buried river channels; others may be the filling of sink holes.

RELATIONS BETWEEN FEATURES OF VALLEY FLOOR

The descriptions just given indicate that the floor of the Shenandoah Valley includes a number of incompatible features, formed under contrasting conditions, and probably widely separated in time. In view of these circumstances, the earlier and simpler conception of the valley floor as a part of the Valley Floor or Harrisburg peneplain tends to disappear. Actually, the widest tracts of valley floor in the Elkton area are the depositional surface of the intermediate gravels, formed at a time by no means as remote as that assumed for the Valley Floor peneplain. The entrenched meanders of the South Fork of the Shenandoah River, commonly assumed to date from the time of formation of the Valley Floor peneplain, likewise seem to have been formed during the closing stages of the period in which the intermediate gravels were deposited. As the residuum that accumulated on the Valley Floor peneplain is now everywhere eroded and overlain unconformably by gravel deposits, it is unlikely that any remnants of the former peneplain surface are still present in the Elkton area. The surface features of the valley are entirely those that were established during the periods of gravel deposition and later.

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[4] Records of Virginia Geological Survey, through courtesy of W. M. McGill, State Geologist, September, 1947.

Both Keith (1894, pp. 374-376) and Fenneman (1938, pp. 247-248) have noted a marked increase in the gradient of the Shenandoah Valley above Front Royal, as compared with its lower or northern section. This gradient is stated to be incompatible with the slope assumed for the Valley Floor peneplain at the time of its completion, and hence to be suggestive of subsequent warping or uplift. Insofar as the Shenandoah Valley in the vicinity of the Elkton area is concerned, the gradients discussed by Keith and Fenneman are not those of the Valley Floor peneplain, but of the gravel deposits. These gradients are in harmony with the regimen of Pleistocene and modern streams, but they are out of harmony with conditions, as they are assumed to have existed, at the time of the formation of the Valley Floor peneplain and its accompanying residuum. A change in gradient has clearly taken place, either by warping or by some other means.

The history of the valley floor before the deposition of the gravels evidently occupied a long period of geologic time, even though the land surface itself was not notably lowered by erosion. First, residuum accumulated slowly as a result of rock weathering, probably in a warm and moist climate, under conditions of peneplanation. Afterwards, deposits such as the ancient gravels were laid down locally on the surface. In Georgia and Alabama, fossil plants in deposits on the valley floor surface indicate that these deposits are of early Tertiary age. The deposits in Virginia may be as ancient, but their age has not been proved. After the deposition of the ancient deposits, both their surface and that of the residuum was eroded before the older gravels were deposited. If the older gravels are Pleistocene, the events which preceded their deposition would seem to require a considerable span of Tertiary time.

According to widely accepted theories of Appalachian evolution, the shaping of the present land forms was accomplished in later Tertiary time. Johnson (1931, pp. 14-21) considers that the Schooley peneplain was not completed until middle Tertiary, and the Harrisburg peneplain not until late Tertiary. Stose, who believes that the Schooley was formed much earlier, now (1940, pp. 461-476) suggests that it was not materially dissected until the Miocene, which would result in assigning about the same date to the completion of the Harrisburg surface. This is not the place to consider the elaborate structure of deduction on which conclusions of these and other authors have been based. The writer can only state that the dates thus assigned for the completion of the Valley Floor or Harrisburg peneplain do not seem sufficiently ancient to account for all the events which appear to have taken place on the valley floor, as indicated by the geological materials observed in the Elkton area.

ACKNOWLEDGMENTS

In this study of the Cenozoic history of the Elkton area, the writer is particularly indebted to his colleague, John Rodgers, who saw in the field many of the features herein described, and who subsequently was a helpful critic of the writer's manuscripts. The writer wishes to express his gratitude to Prof. R. F. Flint for reading and criticizing an early draft of the manuscript of this paper, and for encouraging its preparation for publication. Great aid has also been received from Josiah Bridge, who made helpful suggestions while the writer was working out the Cenozoic sequence of the Elkton area, and who supplied valuable supplemental information based on bauxite investigations of the U. S. Geological Survey. The discovery and proper interpretation of the ancient gravels at the Stanley Mine was worked out in the field jointly by H. D. Miser and the writer. W. M. McGill, State Geologist, has kindly supplied records of water wells from the files of the Virginia Geological Survey.

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LATE PLEISTOCENE AND RECENT CHANGES OF SEA LEVEL ALONG THE COAST OF SANTA BARBARA COUNTY, CALIFORNIA

J. E. UPSON

ABSTRACT. At the mouths of three coastal valleys of Santa Barbara County, California—the Santa Ynez and Santa Maria Valleys and the Goleta Basin—evidence from well borings indicates that Pleistocene streams carved valleys graded to sea level which at that time was at least 200 feet and probably a maximum of about 300 feet below present level. These valleys, now filled with alluvium whose surface is graded about to present sea level, were cut below a marine terrace having a peak shore line about 100 feet above present sea level. Fossils in deposits on the wave-cut bench of the terrace suggest that the terrace is upper Pleistocene in age. The prealluvial valleys are believed to have been cut as a result of lowered sea level accompanying the maximum glaciation of the Wisconsin stage; and the terrace may have been formed during the immediately preceding interglacial stage. The suballuvial valley bottoms are about the same distance below present sea level; and the subsequent relative rise of the sea (about 300 feet) is considered to be a eustatic rise. The previous relative decline (maximum of about 400 feet) is considered to be almost wholly eustatic, although a small part of it may have been due to rise of the land.

Comparison of the longitudinal profile of the suballuvial valley of the Santa Ynez River on the one hand and that of the nearly contiguous Arguello submarine canyon on the other indicates that the two profiles were formed in distinctly different episodes. The formation of the submarine canyon (by whatever process) is thought to be older, and is considered pre-Wisconsin.

INTRODUCTION

PURPOSE OF THE PAPER

DROWNED valleys now more or less wholly filled with alluvium have been described at several places along the California coast. The most outstanding example is San Francisco Bay, but many similar though less conspicuous features have been noted elsewhere. Ellis and Lee (1919) early described alluvium-filled valleys along the streams in San Diego County; and Grant and Gale (1931, p. 65) mention the overdeepened valleys of the main coastal streams. Most writers have ascribed these features specifically or tacitly to an oscillation of the land up and then down with respect to sea level. However, along this and other continental coasts numerous examples of overdeepened stream valleys have been ascribed to eustatic lowering of sea level.

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This paper discusses certain overdeepened valleys along the coast of Santa Barbara County, California, and describes their relation to the youngest and lowest of several marine terraces. It outlines a sequence of geomorphic events during the late Quaternary which is believed to have resulted in large part from eustatic fluctuations of sea level that accompanied the waxing and waning of maximum glaciation during the Wisconsin stage. The sequence is correlated with similar sequences along other parts of the California coast; and a possible bearing on the time of origin of one so-called submarine canyon is suggested.

ACKNOWLEDGMENTS

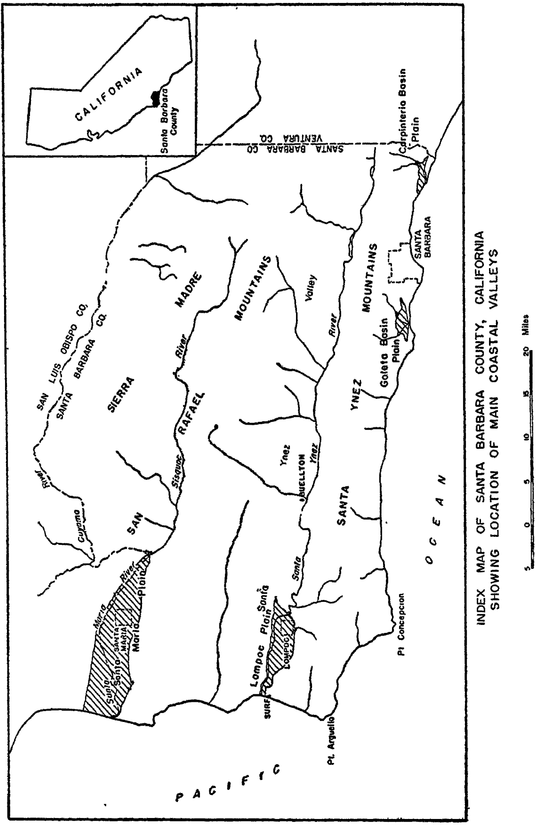
The data and observations here discussed were accumulated during an investigation of the ground-water resources of Santa Barbara County by the U. S. Geological Survey in financial coöperation with the County. The writer wishes to acknowledge the aid and criticism of his colleagues both in the course of the work and in the preparation of this paper. Particular acknowledgment for review, criticism, and timely comment is due Prof. U. S. Grant, of the University of California at Los Angeles, Dr. W. P. Woodring of the U. S. Geological Survey, and Prof. A. O. Woodford of Pomona College. The writer is also indebted for criticism and comment to Prof. R. P. Sharp of the California Institute of Technology.

LOCATION OF AREAS

The areas of principal concern in this paper are the main coastal valleys of Santa Barbara County, California, which is on the Pacific coast about 300 miles south of San Francisco Bay and about 80 miles northwest of Los Angeles (see fig. 1). The county is in the right-angled part of the coast formed by two prominent points close together: Point Arguello, from which the coast trends generally northerly, and Point Concepcion, from which the coast trends easterly. In broad outline the coast is regular and simple, but in detail it has a variety of shore-line features. These range from bold headlands, where the shore line transects mountain ridges, to low plains and swamps at the mouths of the rivers and streams and in certain primarily structural basins along the south coast. Along most of the shore, extensive terrace remnants slope seaward to end at the water or at a narrow beach in abrupt cliffs from 40 feet to several hundred feet high.

The largest coastal valleys are along the lower courses of the two principal streams of the county, the Santa Ynez River

FIGURE I



INDEX MAP OF SANTA BARBARA COUNTY, CALIFORNIA
SHOWING LOCATION OF MAIN COASTAL VALLEYS

Fig. 1. Index map of Santa Barbara County, California, showing location of main coastal valleys.

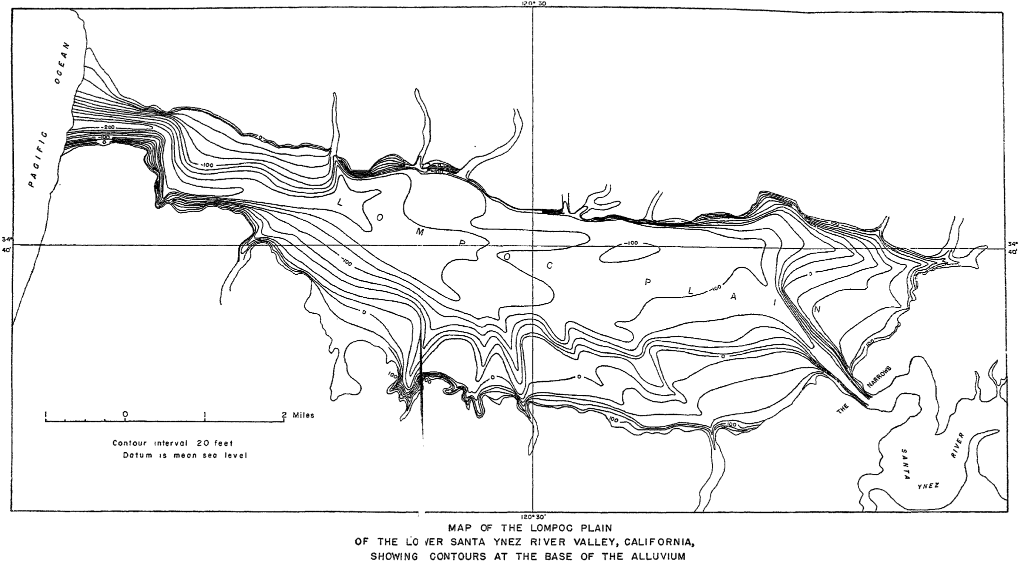


Fig. 2. Map of the Lompoc Plain of the lower Santa Ynez River Valley, California, showing contours at the base of the alluvium.

and the Santa Maria River, which cross rather broad alluvial plains immediately before entering the sea. The plain of the Santa Ynez River, called the Lompoc Plain, extends about 12 miles inland, and that of the Santa Maria River, called the Santa Maria Plain, extends about 20 miles inland. In addition to these larger coastal valleys, two smaller ones, called the Carpinteria and Goleta Basins, lie along the south coast of the County. These are structural basins originated by faulting, and now contain alluvial plains formed by deposits of several short streams.

In the Lompoc Plain, the Santa Maria Plain, and the Goleta Basin, geomorphic and geologic features indicate a like sequence of events explainable in large part by a eustatic decline and rise of sea level. In the Carpinteria Basin not all elements of the sequence are demonstrable, but the history was probably the same. The pertinent features of the first three areas are described in ensuing paragraphs.

RELATIVE CHANGES OF SEA LEVEL IN SANTA BARBARA COUNTY

LOMPOC PLAIN

Geomorphic and geologic features

The Lompoc Plain lies in the lowermost part of the Santa Ynez River Valley, between the Purisima Hills on the north, the Santa Rita Hills on the east, and the foothills of the Santa Ynez Mountains on the south. It is a small alluvial plain which extends landward from the ocean for about 12 miles and ranges in width from half a mile to two miles. It is traversed by the Santa Ynez River, which enters the plain at the southeast corner from a narrow valley, called The Narrows, cut in the consolidated rocks (pl. 1, A), then flows westward along the north side of the plain and enters the ocean at the west end.

The geology of this area has been discussed in some detail in a report on the geology and water resources of the Santa Ynez River Valley (Upson and Thomasson, 1947). Considerable information, especially on the pre-Pleistocene deposits, is contained in the paper by Woodring and others (1943) on the Santa Maria district, which includes part of the Purisima Hills. Only certain salient features are summarized here.

The Lompoc Plain is underlain by alluvium, whose thickness and relationships are significant to the discussion in this paper.

The plain is a series of gently sloping alluvial-fan surfaces, each of which heads in the canyon of its originating river or creek. The surface of the plain thus formed abuts sharply against the sides of the bordering hills and terraces in the fashion characteristic of depositional surfaces. Both the surface of the plain and the Santa Ynez River channel appear to be graded to present sea level, but the river channel is entrenched below the plain surface by a progressively greater amount inland from the shore. Where the river enters the plain, this entrenchment is about 45 feet. Other tributaries on the south also have incised their channels at the margins of the plain in somewhat lesser amounts.

The alluvium occupies a valley which was excavated by the Santa Ynez River in the floor of an older valley now represented by a set of terraces 35 to 90 feet above present sea level and present stream grade. The terraces are capped by alluvial clay, sand, and gravel which rest unconformably on deformed older sediments of Miocene to upper Pleistocene age. Along the southeast and west margins of the plain the alluvium rests on the Monterey shale of Miocene age, but in the middle part of the plain it rests on the relatively unconsolidated marine Careaga sand of upper Pliocene age, the Paso Robles formation of Pliocene and questionable Pleistocene age, and the Orcutt formation of upper Pleistocene age.

These formations have been folded in more or less intermittent deformation with several periods of considerable intensity—the first at the close of Monterey time and the second in the lower Pleistocene after deposition of the Paso Robles formation. Since then some minor warping has taken place, but the Orcutt formation is only slightly tilted, and the deformation appears to be dying out. Within the area of the Lompoc Plain, the deformation developed a syncline whose axis trends southwestward, crossing the Lompoc Plain diagonally. Thus, the Monterey shale at the west end of the Lompoc Plain, which forms the northwest limb of the syncline, moved upward relative to nearby areas during deformation.

Characteristics and origin of the alluvium

In regard to Pleistocene and Recent changes of sea level, the thickness and shape of the alluvium and its relations to underlying formations are critical. About 110 water wells have been drilled below the surface of the Lompoc Plain; nearly all completely penetrate the alluvium and pass a short distance into the underlying formations. In most well logs these underlying formations are readily recognizable; and thus the base of the alluvium is accurately determined at about 100 points. Figure 2 is a contour map of the base of the alluvium based on the altitudes of these points. This map shows that the alluvium fills an old valley which is somewhat narrower than, but roughly corresponds to, the present valley. The old valley extends outward from The Narrows as a narrow, elongate trench, broadens somewhat beneath the central part of the alluvial plain, where it was excavated in the unconsolidated Careaga sand and Paso Robles and Orcutt formations, and then narrows seaward to a fairly narrow, steep-walled canyon in the consolidated Monterey shale bordering the seaward part of the present plain. Smaller tributary canyons can be crudely traced to the mouths of present-day tributaries entering the south side of the valley. The walls of the old valley are relatively steep, and they more or less continue the slopes of the present valley walls against which the alluvial fill was laid. The alluvium ranges in maximum thickness from about 190 feet at the head of the valley to over 200 feet at the seaward end. The base of the alluvium at the present shore line is not precisely known from well records but the gradient of the base up-valley, averaging 10 to 12 feet per mile, is sufficiently well fixed so that it can be projected from the nearest known points the remaining two miles to the shore with fair accuracy. The base of the alluvium and hence the valley floor at the deepest point is determined to be nearly 220 feet below sea level at the present shore line.

The well logs also show many of the characteristics of the alluvial fill itself. The deposits comprise two sharply defined members. The lower member is of coarse gravel which contains rounded cobbles and boulders as well as some sand, and constitutes a continuous, linear body filling the lower part of the old filled valley. It is 60 to 80 feet in maximum thickness and doubtless represents the channel deposit of the old river. The upper member is composed predominantly of fine-grained material—sand, silt, and clay. As indicated by the well logs, these deposits are lenticular and the various classes of deposits interfinger. Sand and silt seem to predominate in the headward and marginal parts of the valley, and clay and silt in the central and seaward parts. No clay beds seem to be continuous across the entire valley, and no marine shells are reported in the well logs. Therefore, it seems that beds of marine or lagoonal clay are rare or absent entirely and essentially all the deposits are stream-laid.

The Lompoc Plain is actually the lowest and youngest of a descending series of valleys. The inter-meander spurs and bordering hills in and immediately upstream from The Narrows have flat summits capped by terrace gravel at three levels (see pl. 1, B) the lowest of which is about 90 feet above present stream grade in The Narrows. The remnants are capped by as much as 30 feet of stream-laid gravel; hence the cut surface is about 60 feet above stream grade. Farther upstream it is locally somewhat less than 60 feet. In the gravels of this terrace several miles upstream, near Buellton, the writer found a fossil thigh bone which has been identified by the late C. W. Gilmore as belonging to the genus *Camelops*, thus dating the deposit as Pleistocene.

Downstream, principally along the north border of the Lompoc Plain, this low stream-cut terrace can be traced almost continuously to the ocean, where it merges with a low wave-cut terrace. This terrace is a sloping bench, in part now blanketed by windblown sand and a thin layer of shore deposits. At most places the deposits are not exposed, but they do crop out at the south side of the mouth of the Lompoc Plain where the road climbs to the town of Surf. Here the deposits are composed of lenses of thin-bedded silt, sand, and gravel, with a layer of coarse gravel at the base which truncates dipping shale beds of Miocene age. The base is about 35 feet above sea level at this place, but away from the river mouth the inland edge of the terrace base is somewhat higher.

From the evidence presented in the foregoing, the following sequence of events in late Quaternary time is outlined. The Santa Ynez River developed a cut terrace graded to a marine terrace roughly about 60 feet above present level. Then base level declined and the river excavated a valley graded to a point at least 220 feet below present sea level. Finally base level rose, causing the Santa Ynez River and tributary streams to aggrade their channels, and alluvium was deposited in that valley, filling it to a level graded about to present sea level.

Except for entrenchment in the upper end of the alluvial plain to a maximum depth of about 45 feet in The Narrows, the deposition of the alluvium is the last event in the area. The alluvial filling occurred after the development and strong dissection of the lowest terrace, which is of Pleistocene age, and thus is late Pleistocene or Recent in age.

SANTA MARIA PLAIN

Geomorphic and geologic features

The Santa Maria Plain is about 15 miles north of the Lompoc Plain. It is similar to, but larger than, the Lompoc Plain, and is much wider at the coast. It is the alluvial plain of the Santa Maria River, which is formed by the junction of the Cuyama and Sisquoc Rivers at the head of the plain. The plain extends inland about 20 miles, and is about 10 miles in maximum width. Its topographic and geologic features have been described in a report on the geology and groundwater resources of the Santa Maria Valley area, by G. F. Worts, Jr. (1947), and also by Woodring and others (1943).

The stratigraphy and geologic history are essentially the same as for the Santa Ynez Valley. The plain is underlain by alluvium which was deposited in a valley cut below a series of river terraces, the lowest of which is about 40 feet above the plain. The alluvium and deposits capping the terraces rest unconformably on the Orcutt formation of upper Pleistocene age, the Paso Robles formation of Pleistocene (?) and Pliocene age, and the Careaga sand of Pliocene age; also on the Foxen mudstone and Sisquoc formation of Pliocene and upper Miocene age, the Monterey shale of Miocene age, and locally on the Franciscan formation of Jurassic (?) age and the Knoxville (?) formation of Upper Jurassic age. Some deformation took place in the early Pliocene, as in the Lompoc area, but the principal Cenozoic deformation took place in the lower or middle Pleistocene after deposition of the Paso Robles formation. Later in the Pleistocene slight warping tilted beds of the Orcutt formation, but since then no appreciable deformation has occurred. Thus, since early Pleistocene time diastrophism has been on the wane.

The Santa Maria Plain, like the Lompoc Plain, is on the site of a major syncline. However, here the axis trends northwesterly and passes seaward beneath the broad west end of the plain. Thus, movement of the valley area near the shore has been, and any tendency for late movement probably would be, consistently downward.

Characteristics and origin of the alluvium

In the Santa Maria Plain the alluvium is less readily distinguished from the underlying deposits than in the Lompoc Plain, but reliable data are at hand from nearly 500 water and oil wells that penetrate the alluvium. These show that the alluvium was deposited in a valley whose lowest point near the ocean is about 230 feet below sea level. The alluvium comprises two members: an upper member of fine sand, silt, and clay that extends inland from the ocean 10 to 12 miles where it grades into the other member of coarse sand and gravel. The sand and gravel member comprises the lower part of the formation near the coast, but the entire thickness inland. The enclosing valley was carved below an older erosional level whose existing terrace remnants are some 40 to 100 feet above the alluvial plain and which is correlated with a low marine terrace along the ocean. The marine terrace is probably the equivalent of the low terrace at the mouth of the Santa Ynez River.

Thus, the sequence of events in the Santa Maria Valley is the same as in the Lompoc area. Streams developed an erosional plain, the youngest in a series of such plains; then with a decline of base level they trenched a valley graded to a sea level at least 230 feet below present sea level; and finally, with a rise of base level, they deposited alluvium in the valley and filled it to a surface graded to present sea level.

GOLETA BASIN PLAIN

Geomorphic and geologic features

The Goleta Basin contains a small alluvial plain about eight miles long from east to west and about three miles in maximum width, which, together with bordering foothills and terraced areas, lies along the south flank of the Santa Ynez Mountains (see pl. 2, A). This alluvial plain is open to the ocean only in a very narrow gap. Along its south (seaward) side, the plain is bordered by a nearly continuous block of deformed Miocene and Pliocene deposits which are truncated by a low marine terrace. Seven sizable streams flow from the mountains and cross the alluvial plain. Near the south side of the plain these streams join into two main channels across the coastal terrace. The channels pass on either side of Mescal "Island"—a small, round, flat-topped remnant of the coastal terrace once isolated by sea water, but now connected to the surrounding terrane by alluvial fill. South of the "Island" the two channels join to form one just before entering the ocean (see pl. 2, B).

The geology of the Goleta Basin has been described in some detail elsewhere (Upson, 1947). The alluvial plain is underlain by alluvium which occupies a valley system excavated by streams below a broad wave-cut plain of which the coastal terrace is the principal existing remnant. This terrace, and higher, older terraces, are cut across deformed rocks which range in age from Eocene to lower Pleistocene. The terrace deposits are considered to be upper Pleistocene in age, and are very little deformed if at all. The terrace deposits, and the alluvium of the plain, rest on unconsolidated marine clay, silt, and sand comprising the Santa Barbara formation of lower Pleistocene age. This formation has been rather strongly deformed—and has been displaced 1,000 or more feet by movement along faults. The Santa Barbara formation, and locally underlying marine clay and silt of upper Pliocene age, in turn rest unconformably on older marine and continental deposits that range in age from Eocene to Miocene. Locally, the Pliocene and Pleistocene sediments are absent and the alluvium rests unconformably on these older formations.

Thus, there have been two periods of deformation in late Tertiary and Pleistocene time—one in the upper Miocene or lower Pliocene following the deposition of the Monterey shale and the other after the deposition of the lower Pleistocene Santa Barbara formation. Earthquakes in the region attest to the continued presence of crustal stresses, but since the post-Santa Barbara deformation, earth movements have apparently been on the decline.

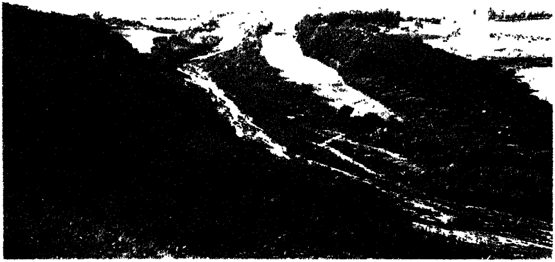
With respect to the current study, the principal result of the post-Santa Barbara deformation was the uplift of the Monterey shale along a reverse fault or fault zone, with upthrow to the south, which about parallels the present shore on a west-trending, curving trace concave to the north. The zone has been called the More Ranch fault by Hill (1932). It lies half a mile to a mile north of the shore and passes along the north side of the coastal terrace, which is cut across the

Monterey shale. This uplifted shale constitutes the south limit of the alluvial plain with the exception of the gap south of Mescal "Island." Thus, the alluvium of the main part of the plain rests on the unconsolidated Santa Barbara formation north of the fault, but in a narrow strip at the south edge of the plain south of the fault, it rests on the consolidated Monterey shale. Furthermore, any late diastrophic movements along the old structural lines would presumably comprise relatively upward and not downward movement of the coastal terrace block along the shore. The significance of these two features is brought out in subsequent paragraphs.

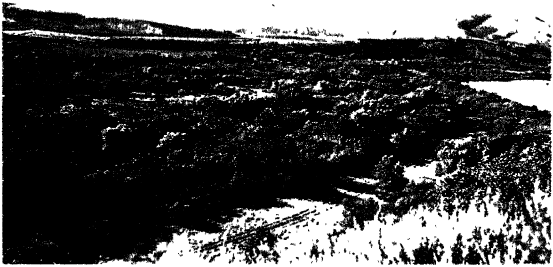
Characteristics and origin of the alluvium

Here, as in the other areas discussed, the thickness and characteristics of the alluvial fill underlying the plain are known primarily from logs of water wells. In the logs the base of the alluvium is fairly readily distinguished from the underlying Santa Barbara formation, which is characterized by distinctive shells of marine mollusks. Sufficient data are not at hand accurately to contour the base of the alluvium, but together with other general information they are adequate to show that the alluvium occupies a valley system which was graded to sea level at a lower position than the present, as in the other areas discussed in this paper. Several logs of oil-prospect holes drilled on a sand spit across the buried outlet disclose a body of sand and gravel overlying the consolidated shale and extending to an approximate maximum depth of 225 feet below sea level. It is inferred that this sand and gravel are the deposits of an older stream system which traversed the Goleta Basin and carved valleys whose currently recognizable low point is about 225 feet below present sea level. Most of the alluvial fill is composed of clay and silt; and any extensive lower member of gravel such as those underlying the Lompoc and Santa Maria plains seems not to exist.

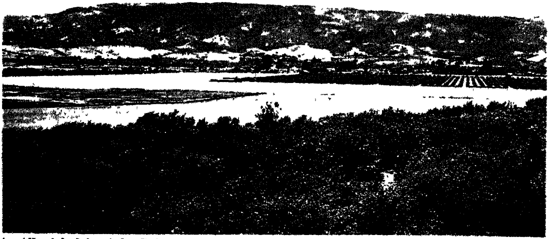
A corresponding sequence of events doubtless also occurred in the Carpinteria Basin, about 20 miles east, but the history there evidently was complicated by continuing movement along coastal faults. Also, the coastal barrier of Tertiary shale is missing along the coast south of the western half of the basin. Deposits which predate the alluvium are not readily distinguishable in well logs from the alluvium; and a prealluvial valley system, though recognizable near the mountains, cannot be traced seaward across the plain.



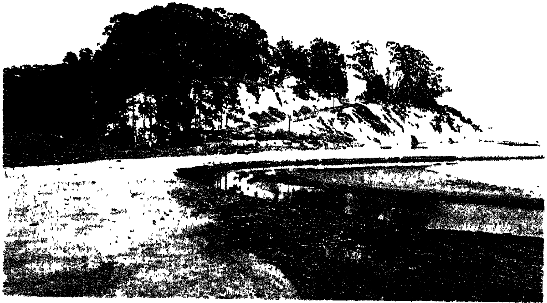
A. View northwest down Santa Ynez River channel at The Narrows Hill at left and terrace remnant at extreme right are composed of Miocene shale. To right of brushy river channel is a plowed field on 15-foot alluvial terrace and farther right the walnut grove is on a 45-foot alluvial terrace continuous with the alluvial plain in the distance.



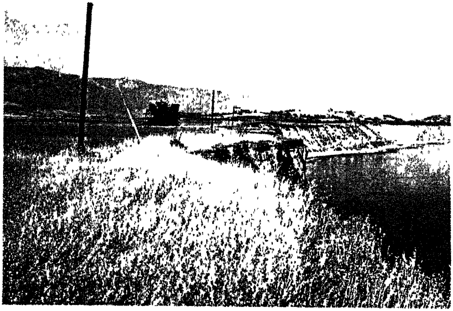
B. View northwest along meandering channel of the Santa Ynez River immediately above The Narrows. Meanders incised in Miocene shale cut by three terraces on right skyline. The lowest terrace is equivalent to the low marine terrace at the coast. Channel is underlain by alluvium to a maximum depth of about 145 feet.



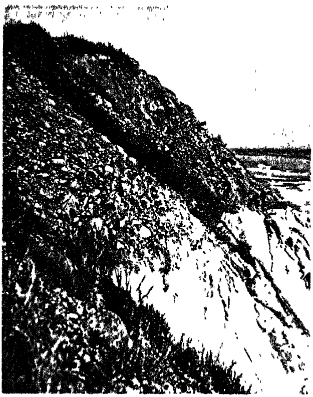
A. Alluvial plain of the Goleta Basin. View north from the coastal terrace; Santa Ynez Mountains in background.



B. View east in outlet of Goleta Basin plain; from a point south of Mescal "Island", and about at the confluence of the two channels flanking the "Island". The cliff nose just left of center is the east side of the gap. It is composed of steeply dipping Miocene shale truncated by marine terrace deposits. Shore deposits and alluvium below the camera site have a maximum thickness of about 200 feet.



A. View east along surface of marine terrace and sea cliff from point about 1 mile east of site of plate 2, B. Veneer of flat-lying deposits truncates north-dipping Pliocene marine beds. Santa Ynez Range and east end of Goleta Basin at left.



B. Terrace deposits on low wave-cut platform truncating steeply dipping Monterey shale near Naples, about 8 miles west of Goleta Basin.

In the Goleta Basin the old, now buried, valleys were cut below an erosional surface, which was probably almost entirely a marine wave-cut platform. This surface is revealed by extensive level surfaces (pl. 3, A) that are seen to level the contorted shale of the coastal block. That these are remnants of a marine platform is indicated by the capping deposits, which consist of a basal layer of coarse gravel and sand locally containing marine shells, and an upper body of silt, sand, and poorly sorted gravel in distinct but irregular lenses (See pl. 3, B). The basal layer of coarse gravel and sand rests with sharp contact on the beveled edges of the underlying deformed beds, and locally these and the boulders have the characteristic rounded pits and depressions bored by pholads. The deposits of the upper unit are composed of poorly sorted, irregularly stratified clayey and silty sand with lenses and zones of sandy gravel. Hence they were probably laid down by streams, although locally there are lenses of fairly well sorted sand that may be marine in origin. At a few localities fossil bones have been found in the upper deposits. In a cut along the Southern Pacific railroad near Capitan, about 12 miles west of the Goleta Basin, track workmen discovered the lower jaw of a mammoth which fortunately was observed in place and removed by Phil C. Orr, Curator of Vertebrate Paleontology and Anthropology of the Santa Barbara Museum of Natural History, who identified the jaw as belonging to *Archidiskodon imperator*. This fossil was found about 15 feet below the top of the deposits in the bottom of a lense of gravel. It is not badly worn and therefore probably was neither transported far before burial, nor eroded from other beds and redeposited. Accordingly, it is thought to represent the age of the deposits in which it was found, which are therefore considered Pleistocene, and together with their lithologic character shows that the deposits are continental.

This terrace extends east and west for several miles from the Goleta Basin, and it is represented by broad, gently sloping or nearly flat remnants. It has been called the Santa Barbara terrace by Grant and Gale (1931, p. 39), who studied fossil shells from the basal deposits; and has been correlated by Davis (1933, p. 1109) with his Dume platform along the coast of the Santa Monica Mountains.

The terrace remnants, where undisturbed, have a seaward slope formed mainly by the surface cover of continental deposits. The wave-cut platform also slopes seaward, but at a lesser angle; and its landward edge, which theoretically should abut against a wave-cut cliff (Davis, 1933) is ordinarily covered by the alluvium. Also, the seaward edge has been eroded back various distances by the Recent sea. Accordingly, the altitude of the surface varies from place to place; and at most places it is impossible to determine the altitude of the peak shore line represented by the platform. At the mouth of the Goleta Slough (pl. 2, B) the wave-cut platforms of the terrace is about 60 feet above sea level. At Goleta Point, about one mile to the southwest, it is about 35 feet. Remnants of this terrace also occur along the inland margins of the Goleta Basin, but there the alluvial cover conceals the actual eroded platform, and its elevation is not known. Farther west, where U. S. Highway 101 crosses the east side of the valley of Tecolote Creek, the eroded platform can be seen in a cut on the old highway some 40 feet above the present highway. The elevation of the platform here, which is about a quarter of a mile from the shore, and a somewhat lesser distance from the inland edge of the terrace, is about 80 feet. The platform is probably somewhat higher inland, but must be appreciably below the surface of the alluvial cover, whose elevation as estimated from the contours of the Goleta quadrangle is about 150 feet. Thus, the maximum height of the wave-cut platform is probably close to 100 feet above sea level.

Thus, the late geomorphic sequence in the Goleta Basin is as follows: waves cut a marine platform 35 to about 100 feet above present level, then streams cut a valley system, with an outlet through the coastal barrier of shale, graded to a position of sea level at least 225 feet below present sea level. Then with a relative rise of the sea this valley was filled with alluvium about to the present sea level. This seems to be unquestionably the last event of the three, and evidently it continued until very recent time because old residents report that 60 years ago tidal water was sufficiently deep near the present middle of the alluvial plain that ocean-going vessels docked at the town situated there.

SUMMARY OF THE GEOMORPHIC SEQUENCE ON THE SANTA BARBARA COUNTY COAST

At three places along the Santa Barbara coast: the lower part of the Santa Ynez River Valley, the lower part of the

Santa Maria River Valley, and the Goleta Basin, waves developed a marine terrace at a maximum present elevation of about 100 feet. With retreat of the sea, streams first deposited alluvial debris in sloping cones and fans on the marine platform, but later entrenched their courses and cut valleys graded to a relatively lowered sea level. At its lowest position that sea level was at least 220 and probably 300 feet below present level. The major streams cut the deepest and largest valleys, but smaller streams also cut steeper, shallower valleys. Subsequently, with a relative rise of sea level, these valleys were filled with alluvium whose upper surface seems to slope evenly about to present sea level.

The cutting of the low marine terrace is the first event in the sequence; and is the least clearly defined. The terrace remnants have been traced along the Santa Barbara coast by reconnaissance only and there may be some errors in correlation. However, it is believed that the low terrace remnants here discussed represent the same event of marine abrasion.

The magnitude of the relative decline of sea level as measured from the approximate elevations of the marine terrace at the shore is 270 feet for the Santa Maria Plain, 255 feet for the Lompoc Plain, and 285 feet for the Goleta Basin. From a peak shore-line level of 100 feet, the maximum declines measured at the shore would be 330, 320, and 325 feet, respectively. As is discussed in the next paragraph, the actual maximum declines are probably somewhat greater.

At the time of maximum decline the shore must have been some distance seaward from the present shore, and streams flowing to it must have had some gradient. Hence the outermost part of the stream courses must have been somewhat lower than the 220- to 230-foot depths at the present shore. As is shown beyond, the Santa Ynez River was probably graded to a point about 300 feet below present sea level. This point corresponds closely with the outer edge of a submerged shelf bordering a large part of the Santa Barbara County coast and also much of the coasts of the Santa Barbara Channel Islands as shown on U. S. Coast and Geodetic Survey Chart 5202. The edge of this shelf is in the neighborhood of the 50-fathom line, or about 300 feet below sea level. This bench has been recognized by other writers on the California coast; and Shepard and Emery (1941, Charts I-IV) show the bench, somewhat discontinuous and of greatly varying width, along the entire California coast. It can hardly represent anything other than a stand of the sea at a maximum of 300 feet below present level; or, for the Santa Barbara coast, about 400 feet below the highest point of the lowest marine terrace now above sea level.

The magnitude of the relative rise to present sea level as measured at the shore is 230 feet for the Santa Maria Plain, 220 feet for the Lompoc Plain, and 225 feet for the Goleta Basin Plain. The agreement strongly suggests that the relative rise was an actual rise of sea level against an essentially stable coast and not a depression of the land. This conclusion is strengthened by the facts brought out in foregoing paragraphs, namely: (1) that differential deformational movements have been slight in late Pleistocene and Recent time in this area, and (2) the directions of any late local differential movements at the mouths of the valleys, if in accord with preexisting structures, would be downward beneath the Santa Maria Plain, but upward beneath the seaward parts of the Lompoc Plain and the Goleta Basin, and hence of no consistent direction for the whole coast. Thus it is concluded that local distrophic movements along existing structures could not have caused the relative rise of sea level.

CORRELATION WITH OTHER AREAS ALONG THE PACIFIC COAST

In the vicinity of Long Beach, California, a similar sequence of events has been described by Poland and others (1945). There streams carved valleys to a depth of about 150 feet at the present shore. The profiles of these valleys project seaward to correspond approximately to a submerged shelf interpreted to be a marine platform, whose inner and outer margins are at depths of about 120 and 240 feet below sea level, respectively. The valleys were evidently cut by streams graded to the sea at the lower level.

Along the coast of San Diego County, Ellis and Lee (1919, pp. 71 and 111-118) describe coastal valleys filled with alluvium whose base extends considerably below sea level. For example, records of wells in the lower parts of the Tia Juana, Mission, and San Diego Valleys (names derived from the respective streams) along the coast between San Diego and the international boundary show that the alluvial fill extends downward to maximum depths of about 100 feet below sea level at the shore. Grant and Hertlein (1944, p. 28) report similar depths of alluvium from a geophysical survey of the Tia Juana Valley. As in the Santa Ynez and Santa Maria Valleys, the alluvium comprises a lower member of coarse gravel and an upper member of finer-grained deposits. Farther north, near the mouth of the San Luis Rey Valley, the base of the alluvium is nearly 200 feet below sea level. The more southern valleys now empty into the ocean in San Diego Bay or Mission Bay, actually some distance back from the open ocean, whereas the San Luis Rey Valley enters the sea at the open coast. Accordingly, it is likely that the depth of fill in that valley is more nearly representative of the true relation between land and sea at the time of valley cutting.

The account of Ellis and Lee (1919, pp. 71, 75, and 110-112) is somewhat obscure as to the precise sequence of events along the coast of San Diego County. However, the latest events postulated by these authors seem to have been relative uplift of the land some 200 feet above present elevation and excavation of stream valleys. This was followed by submergence and filling of the valleys to the present levels of the valley floors. Marine terraces were apparently developed during the earlier uplift. Grant and Hertlein (1944, p. 31) suggest that the valley cutting may have been followed by a rise of sea level to about 100 feet above present level. However, inspection of topographic maps and observations made during a brief trip along the San Diego coast suggest to the writer that the 100-foot stand of the sea preceded the valley cutting, and that the sequence corresponds to that along the Santa Barbara coast.

North of Santa Barbara County similar features are known. Data on the Salinas Valley contained in a report on that area by the Division of Water Resources of the California Department of Public Works (1947) suggest that a similar alluvial valley, graded to a position 200 feet below sea level, may exist in the lower course of the Salinas River. Also, the stream system entering San Francisco Bay once flowed through the Golden Gate, whose maximum depth is about 350 feet below present sea level. Finally, a geomorphic sequence on the Oregon Coast having similar elements has been described by Baldwin (1945). In Oregon the valley of the Coquille River contains a deep fill of alluvium which apparently occupies a narrow valley excavated below a low marine terrace. This terrace has a maximum height of 225 feet above sea level and is capped by late Pleistocene deposits called the "Elk River beds." The sequence here is more complicated than in Santa Barbara County, California, because the low terrace is reported to be tilted, and it also truncates an older estuarine fill called the Coquille formation. The Oregon sequence cannot be surely correlated with the California sequence on the basis of existing data, but the similarity is striking and may be significant.

Thus, along most of the California portion of the Pacific Coast, at least, and possibly also the Oregon portion, streams have apparently carved older valleys below a low marine terrace. These valleys are apparently graded to a relative position of sea level that ranges from 120 to 300 feet below present level. It appears that the larger figure represents the maximum decline of sea level below the present level. These figures, especially the maximum 300 feet, are of the same order as figures assembled by Antevs (1928, p. 82) and more recently by Flint (1947, pp. 446-447) for various parts of the world. Stearns (1935, p. 1934) has presented similar evidence for a 300- to 360-foot stand of the sea below present level on Oahu and other islands of the Hawaiian group.

ORIGIN AND AGE OF THE GEOMORPHIC SEQUENCE

The earliest event of the sequence here described, cutting of the low marine terrace, is considered late Pleistocene in age. The mammoth jaw at El Capitan and the camel femur in the alluvial terrace deposits along the Santa Ynez River suggest this age. Grant and Gale (1931, p. 39) concluded that the marine invertebrates in the lowermost deposits in the terrace near Goleta represent a late Pleistocene age. Grant and Strong (1934) considered that mollusks in tar-impregnated terrace deposits near Carpinteria, probably the same low terrace discussed in this paper, indicate a late Pleistocene age. Finally, Woodring and others (1946) consider all the terraces on the Palos Verdes Hills to be upper Pleistocene. Marine terraces cannot be precisely dated on the basis of contained fossils. Nevertheless, in the Goleta Basin, the low terrace here considered truncates deformed beds of the Santa Barbara formation of which the lower part is considered upper Pliocene by some geologists (Bailey, 1943, p. 1562); but whose upper part is considered lower Pleistocene by most geologists (see Woodring and others, 1940).

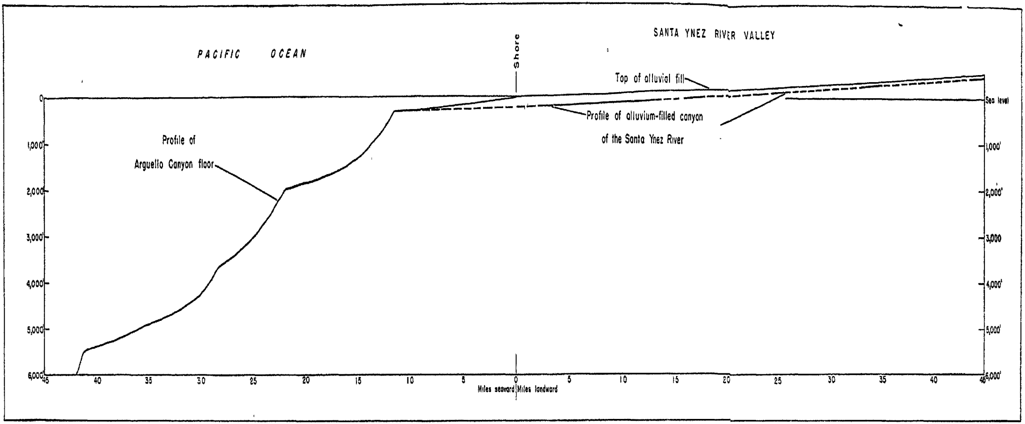
The low terrace must have been developed after the deformation of the Santa Barbara beds, which took place in lower or middle Pleistocene time (Bailey, 1943). Furthermore, it is the lowest and youngest of several terraces along the Santa Barbara coast; description of the higher and older ones of which is beyond the scope of this paper. In order to allow time for a major deformation and the cutting of several older marine terraces in post-lower Pleistocene time, the formation of the low terrace and the initiation of the geomorphic sequence here discussed are considered upper Pleistocene. The geomorphic sequence evidently ends in Recent time with the deposition of alluvium as essentially the latest geologic event of the region. Thus, together with the additional evidence summarized beyond, the intervening valley cutting is considered to have resulted largely from the lowering of sea level that accompanied the maximum glaciation during the Wisconsin stage. The low marine terrace may have been formed in the immediately preceding, or Sangamon, interglacial stage.

In regard to origin, the rise of sea level from the lowest position in the sequence to about the present level is considered eustatic because of its nearly equal magnitude throughout Santa Barbara County, its indifference to the probable direction of local crustal movements, its correspondence in magnitude with similar relative rises of sea level along the rest of the California coast, and its agreement with corresponding relative rises in other parts of the world. Now, if the rise of sea level is a eustatic rise in accord with the waning of a glacial stage, then the immediately preceding lowering of sea level must be in large part eustatic, resulting from a waxing of the same glacial stage and perhaps one or more preceding stages. The writer uses the words "in large part eustatic" because the low marine terrace has not been mapped carefully, its elevation along the coast has not been determined precisely, and it may be slightly deformed at places. A "200-foot" terrace (Putnam, 1942, p. 741) near the west end of Rincon Mountain slopes westward beneath the Carpinteria Basin—apparently tilted. Tilted low marine terraces have been reported at other localities along the California coast. (See Woodring and others, 1946, p. 115 and pl. 26.) However, for many miles along the coast west of the Carpinteria Basin in Santa Barbara County the low terrace certainly is not deformed very much; and it is believed that at least 300 feet, and probably more, of the relative decline of sea level from this terrace was a eustatic lowering of sea level and only a small part of the total difference in elevation can be accounted for by a rise of the land.

The writer does not wish to imply that all terraces older than the 100-foot terrace of this paper are also wholly or even in large part the result of eustatic fluctuations of sea level correlative with glacial stages. At some other localities along the California coast, older terraces are too numerous to be accounted for in this way, for example, on the Palos Verdes Hills, where Woodring, Bramlette, and Kew (1946, pp. 113-116) have described 13 terraces, and on Rincon Mountain, where Putnam (1942) has described 9 elevated terraces. At these localities some of the terraces are tilted, as has been mentioned, and the tilting may be the local expression of more widespread upward crustal movements. Davis (1933) ascribed the preservation of elevated marine terraces of the Santa Monica Mountains largely to uplift, but their origin to eustatic shifts of sea level which he correlated with a hypothetical set of glacial stages. Discussion of the origin and uplift of older terraces of the Santa Barbara coast is beyond the scope of this paper. If, following the lower or middle Pleistocene period of diastrophism, the land continued to rise but at a decreasing rate, the formation of marine terraces developed during the uplift may have been due primarily or entirely to temporary halts during the uplift and may have no genetic relation to glacial stages. On the other hand, to ascribe the origin of the late Pleistocene and Recent cut and fill of the coastal streams, here discussed, to uplift of continental proportions followed by a reversal on the same scale, seems unreasonable.

RELATION TO THE ORIGIN OF SUBMARINE CANYONS

The writer believes that the age and origin of the geomorphic sequence here described has a bearing on the problem



LONGITUDINAL PROFILES OF ARGUELLO CANYON AND OF SANTA YNEZ RIVER CANYON, CALIFORNIA

Fig. 3. Longitudinal profiles of Arguello Canyon and of Santa Ynez River Canyon, California.

of origin of so-called submarine canyons. Off Point Arguello is a minor submarine canyon that heads in four distinct tributaries, one of which seems to be a seaward extension of the Santa Ynez River Valley. This canyon has been called Arguello Canyon by Shepard and Emery (1941), who have described it and others off the California coast.

Figure 3 shows a longitudinal profile of Arguello Canyon above a depth of about 6,000 feet, and a profile of the base of the alluvial fill of the Santa Ynez River. The profile of Arguello Canyon is taken from U. S. Coast and Geodetic Survey chart 5202, dated December 1940. The profile of the base of the alluvium is from well logs and dam-site corings (Upson and Thomasson, 1947, pl. 509). The profiles include distances of about 45 miles seaward and landward, respectively, from the shore. The lines are solid where based on specific data but broken where inferential. The profile of the Santa Ynez River Canyon is well controlled downstream at least to a point about 2 miles from shore. Thence it is projected seaward at about the same slope, as it is considered highly improbable that the gradient would steepen appreciably in that distance. Shepard and Emery (1941, pl. 12), and perhaps others, have drawn comparative longitudinal profiles for Arguello Canyon and the Santa Ynez River and for other seemingly contiguous submarine canyons and land valleys. The profile of Shepard and Emery for Arguello Canyon and the Santa Ynez River was evidently drawn on the surface gradient of the river. It happens that the surface gradient is very little different from the suballuvial profile, but it is believed that the difference is significant geomorphically.

The outstanding features of the profiles here given are (1) the coincidence of the projected suballuvial profile with the outer edge of the submarine bench (now blanketed by shore deposits) at about 300 feet below sea level; and (2) the abrupt steepening of gradient at the head of the profile of Arguello Canyon.

The abrupt change in gradient from the suballuvial Santa Ynez Canyon to Arguello Canyon is a marked physiographic discontinuity. It signifies that the headward part of Arguello Canyon originated in an entirely different geomorphic episode from that in which the prealluvial Santa Ynez River Valley was cut. The discontinuity alone does not indicate which geomorphic episode is the older. If the two profiles were now above sea level, the suggestion would be strong that the lower profile (that of Arguello Canyon) was the younger, and that the abrupt change in gradient was due to the headward cutting of a stream working in an older valley, but graded to a lower position of base level. However, in the existing case, the suballuvial profile of the Santa Ynez River seems clearly to have been graded to the 300-foot submerged bench which apparently truncates the steep head of Arguello Canyon thus suggesting that the steeper profile is the older. Further, if the suballuvial canyon of the Santa Ynez River was cut in Wisconsin time, as argued above, then Arguello Canyon would have had to be cut in pre-Wisconsin time. There seems to have been neither time nor opportunity for subaerial cutting in post-Wisconsin time; and a submarine process, perhaps now operating, would probably be much slower than any subaerial process. Thus, essentially all of the formation of Arguello Canyon, by whatever process, is believed to have taken place in pre-Wisconsin time.

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THE DISTRIBUTION OF ORGANIC MATTER IN SEDIMENTS FROM THE NORTHERN RED SEA

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ABSTRACT. The importance of additional chemical knowledge of the Red Sea bottom sediments is stressed. Methods used for the determination of organic carbon and nitrogen in thirty-eight sediments from the northern Red Sea are described and the results of analysis are given. The distribution of organic matter in these sediments is discussed.

The following points are revealed by the present investigation. 1. The organic matter content of the northern Red Sea sediments is very small. 2. The C:N ratio, which amounts to 5.2, is low. 3. Both the organic matter content and the C:N ratio exhibit regional variations which are ascribed to different environmental conditions of sedimentation. 4. The mean C:N ratio for these sediments is in close agreement with the proportion of these elements in mixed plankton.

INTRODUCTION

OUR knowledge of the chemistry of the Red Sea bottom sediments has been largely derived from the results of the Austrian Expedition on the "Pola" in 1895-96 and 1897-98. Seventeen bottom samples of the "Pola" collection were chemically investigated by Natterer (1898). His analyses included the estimation of carbonate, organic matter, silica, lime, magnesia and some other constituents. Trask (1932) determined the amount of carbonate and nitrogen in ten bottom sediments from the southern Red Sea. Wiseman and Bennett (1940) estimated organic carbon and nitrogen in two bottom sediments collected from the southern Red Sea by the John Murray Expedition. The results of these investigations comprise all knowledge that is available at present of the chemistry of the Red Sea bottom sediments.

Information concerning the abundance and distribution of organic matter in marine sediments is highly desirable. Such information is of special importance to the oil geologist, since it is widely held that the mother substances of petroleum are found in the organic matter of these sediments. It is also of great interest to the marine biologist, since it is believed that the only available source of nutrition for bottom-living organisms inhabiting deep waters is the organic constituents of these sediments.

The direct determination of organic matter in marine sediments is, however, not an easy task. Soil chemists are confronted with the same problem, but as a result of much investigation they arrived at estimating the organic matter content of a soil by multiplying its organic carbon content by the factor 1.724. It is, however, probable that the organic matter in marine sediments is different in its chemical nature and mode of formation from that in soils. Moreover, the constancy of the ratio of organic matter to organic carbon in marine sediments is uncertain. In view of these objections it is preferable to have recourse to some more reliable means of measuring and comparing the organic matter content of marine sediments. As organic carbon and nitrogen are two of the major constituents of organic matter, their determination would provide such means.

It is therefore the object of this paper to record the results of the chemical estimations of organic carbon and nitrogen determined on bottom sediments collected from the northern Red Sea by an Egyptian Expedition, which sailed on the oceanographic boat "Mabahiss" in 1934-35 (Crossland, 1939; Mohamed, 1940), and then to attempt the interpretation of these results.

METHODS

Collection and storage of samples.—Two types of instruments were used for collecting bottom samples on "Mabahiss": the Bigelow sounding rod and the Petersen grab. After collection, each sample was left for some days to dry in air and was then stored; core samples were wrapped in paper and stored in cardboard tubes, while grab samples were stored in glass jars.

Preliminary treatment of samples.—To render the samples absolutely free of salt, 30 grams of each sample were placed in a Jena glass-sintered filter of the finest grade and were repeatedly washed with distilled water until the filtrate no longer gave a precipitate with silver nitrate. Samples were afterwards dried in a hot air-oven regulated at a temperature of 100°C., and then reduced to fine powder, thoroughly mixed and finally kept in previously cleaned, steamed and dried specimen tubes.

Organic carbon determination.—The content of organic carbon was determined by the gravimetric wet combustion method as described by Morgan (1904) and modified by Dixon (1934) so as to be more suitable for the determination of organic carbon in carbonate-rich rocks. This method consists in oxidizing the organic carbon by means of chromic and phosphoric acids and then absorbing and weighing the evolved carbon dioxide.

About 2.5 grams of the dried powdered sample were accurately weighed into a 250 c.cm. round-bottomed Pyrex flask fitted with a two-hole stopper. Through one hole a tap funnel was passed and through the other an upward inclined small condenser to retain most of the steam and acid liberated from the flask. The carbon dioxide present as carbonate was first removed by slowly admitting syrupy phosphoric acid into the decomposition flask from the funnel. When effervescence diminished, the flask was heated briskly until all carbon dioxide present as carbonate was expelled. The condenser was then connected with a train of U-tubes as follows:— (1) a tube containing a mixture of phosphoric and chromic acids; (2) a tube containing glass beads moistened with concentrated sulphuric acid; (3) a tube containing pumice impregnated with anhydrous copper sulphate; (4) a calcium chloride tube; (5) two weighed soda-asbestos tubes; (6) a calcium chloride guard tube; (7) a soda-lime tube which acted as a precautionary trap for any moisture or carbon dioxide leaking from the atmosphere. Before the soda-asbestos tubes were placed in position, a stream of CO_2-free air was passed through the apparatus for 15 minutes to sweep away any carbon dioxide which might be present. The apparatus was then fitted together and 20 c.c. of a 10 per cent solution of chromic anhydride in phosphoric acid were run slowly from the funnel. The flask was heated with a small flame while a gentle stream of CO_2-free air was passed through the apparatus. Heating was continued with a gradually increasing flame for about one and half hours. The flame was then extinguished, but the CO_2-free air current was continued to pass through the apparatus for 15 minutes. Finally, the soda-asbestos tubes were removed and weighed, after standing for half an hour in the balance case, against similar tubes as counterpoise.

The organic carbon content of several samples was determined by duplicate analyses and the results agreed within ± 0.04 per cent. Some typical duplicate results are shown in table 1 following.

TABLE 1

Sample No.	Organic carbon (per cent)	Difference (per cent)
2	0.43, 0.47	0.04
8	0.16, 0.14	0.02
25	0.21, 0.23	0.02
36	0.10, 0.11	0.01

Nitrogen determination. — Nitrogen was determined by the micro-Kjeldahl method. About 2.5 grams of the dried powdered sample were accurately weighed into a micro-Kjeldahl flask; one droplet of mercury was added as a catalyst and 10 – 15 c.c. of nitrogen-free concentrated sulphuric acid were also added very cautiously. Heating was then started with a small flame which was increased gradually and digestion was continued for several hours until decolorization was complete and the precipitated calcium sulphate looked pretty white. The product of oxidation was then filtered into 100 c.cm. measuring flask, the calcium sulphate washed several times with small portions of distilled water and the filtrate made to 100 c.c. The distillation of ammonia and its volumetric determination was subsequently carried out on 10 c.c. portions of this filtrate according to the procedure described by Pregl (1930). Parnas and Wagner's improved micro - Kjeldahl apparatus was used and automatic 10 c.c. micro-burettes graduated to 1/50 c.c. were employed in the volumetric operation.

The nitrogen content of several samples was also determined by duplicate analyses and the agreement between the results was remarkably excellent; these results seldom differed by as much as ± 0.005 per cent.

RESULTS

The results of analysis are tabulated in table 2 below. In the first column the sample number is given, in the second the station number, in the third the geographical position, in the fourth the depth in meters, in the fifth and sixth columns the percentages of organic carbon and nitrogen, and in the seventh the C:N ratio is recorded. Complete list and full particulars of stations are given by Crossland (1939).

Thirty-eight samples were investigated, three of which were procured from the Gulf of Suez, five from the Gulf of Aqaba and the rest from the adjacent regions of the Red Sea proper. In the case of core samples, estimations were made on the upper parts of the cores.

TABLE 2

Results of organic carbon and nitrogen analysis in sediments from the northern Red Sea.

(1) Gulf of Suez

		Position									
Sample No.	Station No.	Latitude °	'	"	Longitude °	'	"	Depth (meters)	Organic carbon (%)	Nitrogen (%)	Ratio C:N
1	1	29	23	30	30	37	00	62	0.35	0.050	7.0
2	2	29	08	15	32	45	30	59	0.45	0.055	8.2
3	3	28	46	45	32	57	15	64	0.38	0.049	7.8

(2) Gulf of Aqaba

4	49	28	27	24	34	30	18	265	0.06	0.024	2.5
5	52 B	28	32	00	34	40	12	1118	0.08	0.023	3.5
6	54	28	40	30	34	42	18	1393	0.11	0.024	4.6
7	55	28	39	12	34	39	12	442	0.12	0.025	4.8
8	57	28	11	54	34	32	12	1128	0.15	0.033	4.5

(3) Red Sea proper

(a) Southern approaches of the Gulf of Suez

9	7	27	30	40	34	03	30	731	0.31	0.049	6.3
10	14	27	19	39	33	48	06	66	0.13	0.029	4.5
11	18	27	19	18	33	49	06	71	0.15	0.042	3.6
12	19	27	17	54	33	50	12	102	0.39	0.052	7.5
13	20	27	22	06	33	46	15	18	0.19	0.029	6.6
14	21 A	27	26	42	33	45	33	73	0.35	0.091	3.8
15	22 C	27	16	24	33	51	18	102	0.38	0.082	4.6
16	23 C	27	18	24	33	52	36	93	0.11	0.025	4.4
17	26	27	27	48	33	54	18	236	0.11	0.041	2.7
18	28 A	27	29	00	33	53	24	118	0.14	0.052	2.7
19	32 F	27	16	30	33	52	12	102	0.24	0.058	4.1
20	34 A	27	23	24	33	56	30	124	0.11	0.030	3.7
21	34 E	27	23	42	33	57	12	88	0.07	0.028	2.5
22	34 H	27	24	03	33	57	48	144	0.09	0.024	3.8
23	34 L	27	23	30	33	57	42	440	0.30	0.072	4.2
24	37	27	24	48	33	59	15	366	0.19	0.039	4.9
25	40	27	34	36	34	00	51	941	0.22	0.086	2.6

(b) Southern approaches of the Gulf of Aqaba

Sample No.	Station No.	Position						Depth (meters)	Organic carbon (%)	Nitrogen (%)	Ratio C:N
		Latitude			Longitude						
		°	′	″	°	′	″				
26	41 H	27	54	57	34	37	38	24	0.09	0.044	2.0
27	42	27	49	30	34	39	30	741	0.12	0.032	3.8
28	43	27	52	00	34	27	24	1147	0.21	0.041	5.1
29	58	27	51	30	34	38	12	596	0.20	0.055	3.6

(c) Open region of Red Sea

30	45 A	27	39	30	34	23	30	960	0.09	0.016	5.6
31	67	27	08	42	34	10	00	814	0.18	0.031	5.8
32	76	26	57	36	34	23	36	951	0.15	0.030	5.0
33	77	27	05	24	34	37	00	1274	0.13	0.023	5.7
34	78	27	13	21	34	50	12	1148	0.11	0.019	5.8
35	81 B	26	48	30	35	40	36	873	0.15	0.028	5.4
36	82 A	26	38	00	35	21	30	834	0.11	0.018	6.1
37	83	26	34	33	35	03	45	1234	0.29	0.047	6.2
38	86	26	13	15	34	36	12	667	0.13	0.024	5.4

DISCUSSION

Before discussing the above results a short survey will be given of the chief factors which determine the content of organic matter in marine sediments. These factors include the supply of organic matter, its transportation by currents, its rate of sedimentation and the extent of its decomposition as well as the configuration of the sea bottom and the texture of the sediments.

In deep and open waters the supply of organic matter depends mainly on the abundance of phytoplankton, while in coastal waters sedentary vegetation is, as a rule, the chief source of organic matter. The plankton productivity of an area depends on a number of factors such as light, temperature, turbulence, the concentration of nitrate and phosphate and other nutrients, and the presence of iron and some other minor constituents in sea water. Consequently, the supply of organic matter will be abundant if the conditions of plankton productivity are favorable. But it does not follow that the underlying sediments will be rich in organic matter, since the transporting influence of the currents may carry organic residues considerable distances. Furthermore, the supply of organic matter may be identical in two areas, yet their sediments may differ in their content of organic matter if there are different rates of deposition of inorganic detritus.

The amount of decomposition which takes place in organic debris before and after sedimentation exerts also a great influence. Before sedimentation, the amount of decomposition depends on the quantity of dissolved oxygen in the water; the smaller the supply of oxygen, the richer is the organic matter content of sediments. It also depends on the depth of water through which the organic debris may sink; the deeper the water, the longer the interval required to reach the bottom and the greater is the amount of decomposition. After deposition, organic debris will continue to decompose through the agency of microörganisms. The extent of this decomposition is largely dependent on the oxygen content of bottom water.

Finally, the configuration of the sea bottom and the texture of the sediments were found to have a profound influence on the organic matter content of sediments. Trask (1932) found that sediments in depressions contain more organic matter than do those on ridges and slopes. He also found that the organic matter of sediments increases as their texture becomes finer.

As to the topography of the area investigated, it is to be noted that the differences between the Gulf of Suez and the Gulf of Aqaba, into which the Red Sea bifurcates at its northern end, are striking. The Gulf of Suez is a shallow flat-bottomed shelf with an average depth of 50 meters, while the Gulf of Aqaba is a deep trough with depths up to 1830 meters in the middle of the Gulf. Moreover, the Gulf of Suez descends at its mouth abruptly to a depth of over 300 meters in the Strait of Jubal, while the Gulf of Aqaba is separated from the deep parts of the Red Sea by a shallow sill in the Strait of Tiran. The Red Sea itself, which is of considerable depth, is studded in places with rocky islets and is characterized by the irregularity of the bottom and the occurrence of submarine hills. No rivers flow into the Red Sea, and the northern part is rainless.

In the light of all these considerations the distribution of organic matter in sediments from the northern Red Sea will be discussed. The organic carbon content as well as the nitrogen content will be used as the basis for describing and comparing the organic matter content of the investigated sediments.

Distribution of organic matter.—The most striking result of the present investigation is that the organic matter content of sediments from the northern Red Sea is very low compared with that of other marine sediments; the organic carbon content ranges from 0.06 to 0.45 per cent and the nitrogen content from 0.016 to 0.091 per cent. Granted that the organic matter content of sediments is an index of productivity, these figures would clearly indicate that organic production in the northern part of the Red Sea is very poor on the whole.

Regarding the distribution of organic matter in the different regions of the investigated area it is found that the organic matter content of sediments from the shallow Gulf of Suez is relatively high; the organic carbon content is greater than 0.30 per cent and the nitrogen content is greater than 0.045 per cent. No quantitative plankton observations were made, but the fact that the transparency of water was lowest in the Gulf of Suez (Crossland, 1939) indicates that the subsurface layer must have been relatively richer in plankton than the other investigated regions. This explains why the sediments in this gulf are relatively richer in organic matter.

In contrast with the Gulf of Suez the organic matter content of sediments in the southern half of the Gulf of Aqaba is very low; the organic carbon content is 0.15 per cent or less and nitrogen content is smaller than 0.035 per cent. As to the cause of the poverty in organic matter of these sediments, it is suggested that it is due to two main factors. One of these factors is the scarcity of plankton, inferred from the high transparency of water (Crossland, 1939), which would result in a low supply of organic matter. The second factor is the rapid rate of decomposition of organic residues favored by the special environmental conditions in this region. Of these conditions the high oxygen content of water, which was greater than 60 per cent of saturation at any depth, the high temperature of the water, which was higher than 21.1°C. at any level, and the great depth of water are the most important.

The distribution of organic matter in sediments from the Red Sea proper presents a complicated picture. Thus, in the southern approaches of the Gulf of Suez the organic matter content of sediments is relatively high at several places and comparable with that of the Gulf of Suez sediments. The nitrogen content is greater than 0.045 per cent in sediments from stations 19, 21A, 22C, 28A, 32F and 34L situated west of Shadwan Island, which lies at the southern end of the Gulf of Suez, and at stations 7 and 40 situated east and north of Shadwan Island respectively; the organic carbon content is greater than 0.30 per cent at a good number of these stations. The relative richness in organic matter of these sediments is possibly due to the transporting influence of the outflowing bottom current from the Gulf of Suez (Mohamed, 1940) which would carry the organic remains from the gulf and deposit them west and east of Shadwan Island as it slows down. However, at stations 14, 18, 20, 23C, 26 and 37 the organic matter content of sediments is rather low. It would seem that other factors such as the configuration of the bottom at these stations or the texture of sediments have a greater influence than that of the outflowing current from the Gulf of Suez. The influence of the configuration of the bottom on the organic matter content of sediments in this region is well illustrated by our findings at station 34 which lies over a submarine hill west of Shadwan Island. At the foot of the hill (station 34L) the organic carbon content is 0.30 per cent and the nitrogen content is 0.072 per cent, while at the other stations which lie on the slopes of the hill (stations 34A, 34E and 34H) the organic carbon content is less than 0.15 per cent and the nitrogen content is less than 0.035 per cent, i.e., less than half the amount found at the foot of the hill.

On the other hand, the amount of organic carbon found in sediments from the southern approaches of the Gulf of Aqaba is low and varies from 0.09 to 0.21 per cent, but the amount of nitrogen, which varies from 0.032 to 0.055 per cent, is somewhat high. The cause of this high nitrogen content is not very evident, although it is quite possible that some local conditions tend to render these sediments specially richer in nitrogenous constituents. Alternatively, it is probable that the environmental conditions in this region are such that the non-nitrogenous constituents of these sediments would decompose more readily than the nitrogenous constituents.

In the open region of the Red Sea proper the organic matter content of sediments is generally low and varies within narrow limits; the organic carbon content ranges chiefly from 0.11 to 0.19 per cent and the nitrogen content from 0.016 to 0.031 per cent. However, the sample collected from station 83 has a somewhat higher content of organic carbon and nitrogen. It is interesting to note that upwelling of water, revealed by oceanographical observations, was taking place in the vicinity of this station. Consequently, supplies of nutrients brought to the surface would result in appreciable plankton production. Hence, the organic matter content of the bottom sample from this station was high.

From the foregoing it appears that the environmental conditions of deposition in the investigated region of the Red Sea proper, excepting the southern approaches of the Gulf of Suez, differ but little from those in the southern half of the Gulf of Aqaba and that poor organic production and rapid rate of decomposition are two factors dominant in determining the organic matter content of sediments in this region.

The carbon-nitrogen ratio.—The investigation of the carbon-nitrogen relation in marine sediments attracted the attention of many workers. Trask (1932) found that the average C:N ratio for 108 marine sediments collected from the Channey Islands area was 8.4, with variations from 5.3 to 13.5. Gripenberg (1934) found that the mean C:N ratio for 81 sediments from the Baltic Sea was 10. Wiseman and Bennett (1940) found that the average C:N ratio for 44 sediments from the Arabian Sea was 14.4, with variations from 4.9 to 34.2.

To investigate the carbon-nitrogen relation in the northern Red Sea sediments, the C:N ratio was calculated and the results are entered in table 2. The average C:N ratio is found to be 5.2, with variations from 2.0 to 8.2. It is evident that this ratio is lower than the corresponding ratio for sediments from other oceanic regions.

From the standpoint of the C:N ratio regional differences are found. Thus, the mean C:N ratio for the Gulf of Suez is 7.7, for the Red Sea proper is 4.6 and for the Gulf of Aqaba is 4.0. These differences would indicate that the conditions influencing the chemical make-up of organic matter deposited in each region are different. They further indicate that the Gulf of Suez sediments which are relatively richer in organic matter have a higher mean C:N ratio than that of sediments from the other two regions which are poor in organic matter.

Granted that the origin of organic matter in marine sediments is due chiefly to the deposition of organic residues of plankton, it is pertinent to enquire how far the mean C:N ratio for the northern Red Sea sediments would agree with the proportion of these elements in the organic content of living plankton. The C:N ratio for mixed plankton as found by Fleming (1940) is 41:7.2 or 5.7:1 which is also in close agreement with the proportion of these elements in sea water. It is found that this ratio is in good agreement with the mean C:N ratio for the northern Red Sea sediments. It is even in perfect agreement with the mean C:N ratio for sediments procured from the stations situated in the open region of the Red Sea proper (stations 45, 67, 77, 78, 81B, 82A, 83 and 86); the mean C:N ratio for these sediments is also 5.7. In view of this agreement our results would seem to suggest that the C:N ratio in marine sediments might come to reflect the proportion of these elements in the organic content of living plankton and in sea water, at least as far as the open sea is concerned. Such a generalization would be of great interest.

SUMMARY

(1) The amount of organic carbon and nitrogen has been determined on bottom sediments from 38 stations in the northern Red Sea. It is found that the content of organic matter in these sediments is very low compared with that of other sediments.

(2) The distribution of organic matter in sediments from the various regions of the northern Red Sea has been discussed. It is found that sediments from the shallow Gulf of Suez are richer in organic matter than those from the Gulf of Aqaba or the Red Sea proper. The causes of these variations have been discussed.

(3) The mean C:N ratio of 38 sediments from the northern Red Sea has been found to be 5.2 which is lower than the corresponding ratio for other marine sediments. Regional variations in the C:N ratio are also found and it is suggested that they are due to different environmental conditions of deposition.

(4) The mean C:N ratio for bottom sediments from the northern Red Sea has been found to be in good accord with the proportion of these elements in mixed plankton.

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MOOREOCRINUS BOWSHERI, NEW SPECIES FROM THE CHESTER SERIES OF NORTHEASTERN OKLAHOMA

HARRELL L. STRIMPLE

ABSTRACT. *Mooreocrinus bowsheri*, new species, is described with a discussion of possible affinity with the rare eleutherozoic *Cryphiocrinus*.

INTRODUCTION

THE presence of a pseudo-Agassizocrinus fauna in the Chester formations of northeastern Oklahoma has been known to the writer for many years. It was not until Arthur Bowsher* discovered a specimen near Cedar Crest Lake that the true nature of the forms became apparent. Only the anterior side of a dorsal cup with some of the arms were exposed firmly imbedded in a hard, blue-gray limestone. Intensive preparation freed the posterior side of the dorsal cup and the lowermost ossicles of the arms. Affinity obviously was within the Cromyocrinidae rather than the Agassizocrinidae. Much information is needed before we will fully understand the genetic lines involved among forms at present assigned to the Cromyocrinidae. This species is closely similar to *Cryphiocrinus* Kirk but the writer has chosen to place it under *Mooreocrinus* Wright and Strimple because of the arm structure. General calyx outline and structure is that of *Cryphiocrinus* but in the basal area fundamental differences exist. A small stem is present, and a comparatively large, deep, basal invagination. There is no indication of a centrale; however, the IBB circlet is almost entirely covered by the proximal columnal so that definite observation is difficult.

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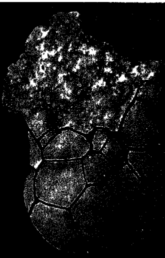
EXPLANATION OF PLATE 1

Mooreocrinus bowsheri, new species.

Figs. 1-3. Posterior, basal, and anterior views of holotype. Enlarged approximately two times natural size.

Hybocrinus crinerensis, new species.

Fig. 4, anterior view of paratype; Fig. 5, posterior view of paratype; Fig. 6, summit view of paratype; Fig. 7, posterior view of holotype; Fig. 8, anterior view of holotype. All figures natural photographs x 2¼.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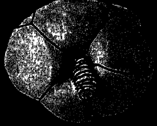
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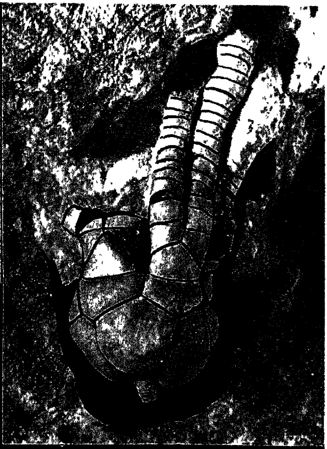
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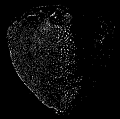
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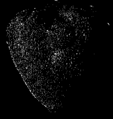
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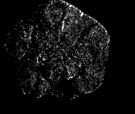
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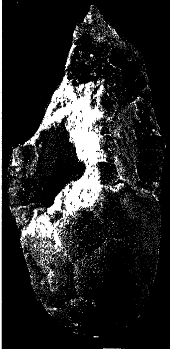
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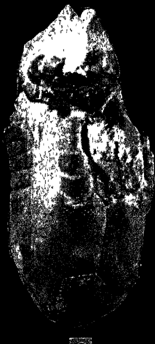
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Through loss of the stem, reduction of basal concavity and antrophy of IBB elements (followed by formation of a centrale) such a form as *Cryphiocrinus girtyi* Kirk could easily be derived. Kirk noted the presence of a minute stem cicatrix approximately .5 millimeters in diameter in the type of *C. rotundus* but considered the stem to be already detached at this stage. In an older paratype of the species a centrale covered the IBB. This is probably a correct assumption because similar occurrence has been noted among the agassizocrinids where the stem impression is present in young specimens, but entirely obliterated in older specimens.

It is recognized that normal evolutionary trends among these late inadunates is for a conical cup to become rounded, the base flat, with a deep basal invagination appearing in advanced stages. It is also understood we are involved in a highly specialized trend in this instance where a transition from statozoic to eleutherozoic habit is probably involved. Following such reasoning the basal area of *M. bowsheri* is only superficially advanced. Presence of a column is indicative of the primitive nature of the form in such a trend and is supported by the primitive three plate arrangement of the posterior interradius. In *C. girtyi* the RX is a modest element and has apparently migrated entirely out of the dorsal cup in *C. rotundus.*

Brief comparison with members of the Cromyocrinidae discloses the following: *Ureocrinus* Wright and Strimple, *Ulocrinus* Miller and Gurley, *Cromyocrinus* Trautschold, and *Tyrieocrinus* Wright all have broad, convex IBB circlets. *Mooreocrinus* Wright and Strimple and *Parulocrinus* Moore and Plummer normally have a slight basal concavity but retain strong, wide IBB circlets. *Ureocrinus*, *Cromyocrinus* and *Mooreocrinus* are the only allied genera having the primitive three plate plan to the posterior interradius, however, in *Ureocrinus* considerable variation is known to exist. *Ureocrinus* and *Cromyocrinus* have five uniserial arms and *Mooreocrinus* has ten uniserial arms. In *Ulocrinus* and *Parulocrinus* only *RA* and anal X are retained within the cup and in *Tyrieocrinus* a single anal plate is present.

The name *Mooreocrinus bowsheri*, new species, is proposed in slight acknowledgement of many contributions of Carboniferous crinoid specimens made by Arthur Bowsher.

DESCRIPTIONS OF SPECIES

MOOREOCRINUS, Wright and Strimple

Mooreocrinus bowsheri, new species

Plate 1, Figs. 1-3

Description.—Dorsal cup globose, constricted at the base of the arms and having a well defined, deep basal invagination. IBB circlet is almost entirely obscured by proximal columnal and apparently forms a flat surface at the bottom of the basal concavity. BB are large elements, proximal portions forming sides of basal invagination and curving sharply upward to form the erect sides of the dorsal cup. Post. B has seven facets, supporting small anal X above and large RA to the upper right. RA in turn supports a small RX above. RR are relatively small, five-sided plates and the uppermost surface curves inward so that meeting with PBrBr is not flush. Sutures between RR and BB are impressed. PBr are axillary and are slightly longer than wide. There are ten uniserial arms composed of moderately thick, rounded ossicles. There is no evidence of ornamentation though the surface of plates are rough appearing. Proximal columnals are thin, round, rather small and taper rapidly.

Measurements.—

	Holotype
Greatest width of cup	12.0 mm.
Height of cup	8.3 mm.
Width of proximal columnal	1.7 mm.

Remarks.—The pronounced basal invagination, presence of a small column and the somewhat elongate dorsal cup serve to readily distinguish *M. bowsheri* from other described species.

The species does not agree in some respects with the diagnostic characteristics of *Mooreocrinus* which form has a rather broad dorsal cup, with a shallow basal concavity and a wide IBB circlet. The arms of *Cryphiocrinus* are not known which prevents complete consideration of affinity with that genus.

Occurrence and horizon.—Unnamed formation (sometimes referred to as the Mayes formation), upper Chester series, Mississippian (lower Carboniferous); Cedar Crest Lake, Center north line, Sec. 20, T19N., R20E., northeastern Oklahoma.

Holotype.—To be deposited in the U. S. National Museum.

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BARTLESVILLE, OKLAHOMA

HYBOCRINUS CRINERENSIS, NEW SPECIES FROM THE ORDOVICIAN OF OKLAHOMA

HARRELL L. STRIMPLE AND WILLIAM T. WATKINS

ABSTRACT. *Hybocrinus crinerensis*, new species from the Bromide formation of the Criner Hills near Ardmore, Oklahoma, provides some additional information and increased geologic and geographic range for this unique genus.

TWO magnificent specimens from the collection of Wm. T. Watkins of San Antonio, Texas, are presented as *Hybocrinus crinerensis*, new species. They were collected several years ago by Watkins at Rock Crossing, Criner Hills, southwest of Ardmore, Oklahoma, from the Bromide formation (Ordovician), and consist of a splendid dorsal cup with tegminal structure in excellent preservation, and a larger cup with arms attached which is the holotype.

HYBOCRINUS CRINERENSIS, new species

Plate 1, Figs. 4-8

Dorsal cup conical shaped, ratio of height to maximum breadth about 2:2.2. BB are large, spread rapidly from the small stem impression. RR are large elongate elements with the exception of r. post. R which is small and rests on the right shoulder of the large RA. Articular facets are small, constricted, and are devoid of articular ridges and grooves. Anal X is small, not elevated, and rests against the l. post. R, with support to the right by the left shoulder of RA and lateral side of r. post. R. There is an opening behind the anal X bounded by the posterior oral. No evidence of minute covering plates for the anal aperture has been found in either specimen, however, they might not be preserved.

There are five orals, the larger being posterior in position and attaining a greater height than approached by the others. A pronounced madreporite is present on post. O. Ambulacral grooves rest on the surface of the tegmen and converge to the center. Small covering plates are present but are mostly disturbed from their normal positions.

The five rami appear weak but seem to have attained considerable length. They are composed of rounded elements with sutures impressed. No pinnular structure has been observed.

Stem cicatrix is round, devoid of crenulations, and is pierced by a small pentalobate canal.

All plates are strongly porous, including the orals and rami.

Measurements.—

	Holotype	Paratype
Maximum height of dorsal cup	13.0 mm.	9.6 mm.
Maximum width of dorsal cup	12.2 mm.	8.9 mm.
Length of r. ant. B	11.9 mm.	4.2 mm.
Width of r. ant. B	4.8 mm.	3.5 mm.
Length of suture between BB	4.5 mm.	2.9 mm.
Length of ant. R (to proximal edge of articular facet)	5.5 mm.	5.0 mm.
Width of ant. R	5.8 mm.	5.5 mm.

Measurements.—continued

	Holotype	Paratype
Length of anal X	3.6 mm.	2.0 mm.
Length of RA	6.0 mm.	4.6 mm.
Width of RA	7.4 mm.	6.1 mm.
Diameter of stem impression	2.4 mm.	1.3 mm.
Width of arm articular facet	2.6 mm.	2.0 mm.
Length of arms (Maximum preserved)	16.3 mm.	—

Remarks.—*H. crinerensis* is close to the genotype species. *H. conicus* Billings, differing mainly in the following respects: more obliquely placed anal X; a broader, fuller dorsal cup; lower placement of anal opening; less protrusion of r. post. R. The right posterior portion of *H. conicus* is very distended, causing an asymmetrical appearance to the cup, particularly when viewed from the summit. Although the r. post. R of *H. crinerensis* has a high position in the cup, and there is a mild protrusion, no decided appearance of irregularity is attained.

A peculiar twist to the dorsal cup in the basal area is of conjectural value. When the cup is viewed from the posterior side, the basal elements twist from left to right and slightly forward. This same development is found in *H. conicus* and indicates a directional leaning. Possibly the lack of pinnular appendages forced the forms to adopt a permanent leaning attitude into the current to augment their food gathering powers. The writer has noticed this slight basal twist in a later (Pennsylvanian) non-pinnular form, *Allagecrinus*, wherein the column is normally curved in the proximal region and sporadic arm development is prevalent.

Occurrence and horizon.—Bromide formation, Ordovician; Rock Crossing, Criner Hills, southwest of Ardmore, Oklahoma.

Types.—To be deposited in the U. S. National Museum.

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BARTLESVILLE, OKLAHOMA

harmonize modern taxonomic theory with paleontological practice.

FUNCTIONS OF TYPES

Type specimens in ordinary paleontological practice have served as 1) bases for definitions, 2) as standards for comparison, and 3) as fixed biological points in variable populations to which binomial names are attached. According to Simpson, "the modern conception of taxonomy as involving the inference of population characters from samples makes it impossible for the same items properly to serve all three of these purposes. . . . The newer theory of taxonomy as a system of group concepts based on inferences about populations from samples . . . is decisively incompatible with this use of 'types'." (*Op. cit.*, p. 413.)

The three-fold function of types still generally in use in paleontology is a natural heritage of the "type" concept in neontology, only recently generally abandoned by students of living organisms. It is inevitable that the paleontologists, following the lead of zoölogists and botanists, will eventually learn to view their collections of specimens as *samples of populations*. As such, collections cannot individually define the characters of species. They can serve only as useful and valid measures of the particular populations to which they belong, provided, of course, that they are properly analyzed.

Species and subspecies are real in nature. They are groups of individuals that are closely related and that vary between limits, however vague and arbitrary.[1] Our concepts of species and subspecies are, of course, subjective and probably never can be more than rough approximations of the true species in nature. Nevertheless, one of the main objectives, perhaps the chief goal of the taxonomist, is to learn as much as possible about the subspecies or species, and every effort should be made to accumulate accurate and adequate information about their characters. Most forms are far more variable than commonly is recognized. An elementary statistical analysis generally is needed for a valid quantitative statement

[1] For example, it is probable that evolution in time is generally or invariably a continuous process. Limits of species are drawn at convenient gaps in the fossil record. There is no reason to believe, however, that these gaps are universal, and annectent forms found after species boundaries are arbitrarily set must be dealt with just as arbitrarily.

of that variability. Likewise, two subspecies of a given species may be clearly and regularly separable only by means of appropriate quantitative methods.

THE HYPODIGM

Traditionally, the original or type collection is the basis for defining and limiting species. Individual variants which fall well outside the artificial limits set by the type specimens necessarily fall in some other category (species, subspecies, or "variety"). Under these circumstances the types *are* the taxonomic species. Clearly, species established on these premises have very little in common with species in nature. Furthermore, there is no provision in this method for accumulating new knowledge about natural species.

Simpson (1940) has shown by means of his *hypodigm* concept, how collections of specimens can provide significant data on natural species and this knowledge can be cumulative as new specimens are acquired. The hypodigm is the *population sample* by which characters of the entire population are assessed. The composition of the hypodigm changes as new specimens are added to it, and as original types or other specimens are removed as required by revision. At the time of description of a species (or subspecies) the hypodigm consists only of the type lot.[2] Subsequent critical studies of the species based on additional specimens should add to knowledge of the species *by increasing the size of the hypodigm.* The hypodigm then is really a statistical sample. As this sample increases in size by acquisition of new collections (by various workers at various times) the appraisal of the whole population, if *uniformly based on quantitative methods,* gradually approaches the facts in nature. The taxonomic species then will be objective and will resemble natural species composed of one or several whole populations.

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[2] It appears to me that the type lot includes *all* of the specimens referred to the new species, subspecies, or "variety" by the describer in the original publication. If no holotype is selected all of the specimens are syntypes (cotypes). If a holotype is designated all of the remaining specimens are paratypes. A limited survey of usage, however, shows that some paleontologists adhere to the view that paratypes or syntypes must also be separately designated. According to this view specimens unspecified in the original publication are not types at all. I know of no authority for this, and it seems to me basically unsound. Even those specimens not referred to in the description and not critically studied enter into the author's impression of his species and influence the result. Hence, they form part of the type lot.

The possibilities inherit in this concept are overwhelming in implication, and should lead to complete revision in paleontological thought and procedure.

FOSSIL POPULATIONS

There are practical difficulties and limitations in use of the hypodigm which I believe require careful consideration. The function of the hypodigm is to enable us to learn as much about natural populations as possible through the use of samples from those respective populations. Unfortunately, it is not at all clear how a fossil population may be defined or recognized.

A brief survey of neontological literature indicates that a population, amongst living organisms, is an effectively interbreeding aggregate which forms a gene pool more or less apart from other closely related aggregates. Through isolation, which may be of several degrees and kinds, a population may acquire distinctive gene properties. Under these conditions a subspecies may arise, being co-extensive with one or more populations. Further evolution may produce separate species, which generally comprise a large, but variable number of populations. In space, populations tend to be segregated geographically, or ecologically. In time, however, they certainly must occupy points in a continuous process of organic change. For this reason the vertical boundaries of fossil populations are vague in the extreme. Elsewhere (Newell, 1947, p. 163) I have tried to define (living and fossil) populations in time. "We may speak of past or future populations, or of successive populations. Thus, a population does not have time duration. A subspecies certainly does have time duration (because its vertical limits may be arbitrarily fixed), hence, a subspecies is composed of a succession of gradually changing populations."

It will be seen that only under closely limiting conditions can a hypodigm of fossils be restricted to specimens from a single population. Approximate contemporaneity in a collection of fossils cannot always be proved, and it is probable that most accumulations of fossil skeletons represent an appreciable span of time. At best then all of the specimens in a hypodigm clearly should be closely related, and should exhibit approximately the random variation characteristic of single or very closely related populations. If the hypodigm is not critically assembled, significant errors through contamination will enter into the cumulative results. By definition "every specimen definitely referred (in publication) to a species is, or has been, part of its hypodigm" (Simpson, *Op. cit.*, p. 419). That is, every specimen studied and definitely referred by an author to a species is part of his statistical sample for that species, *i. e.*, is part of *his* hypodigm. Other authors, using additional material, will of course, have somewhat different hypodigms. Successive descriptions will progressively lead to refinement of elimination of spurious data, and there will be an overall tendency for the hypodigms to become more comprehensive and taxonomic boundaries will be extended with discovery of new specimens. In this way knowledge of variation in natural groups is cumulative rather than static as in the "type" method of comparisons with fixed standards.

Homogeneity of the sample is, of course, one of the primary objectives of the hypodigm. It is not implied that a fossil species or subspecies should be co-extensive with single populations. The hypodigms, even of living forms, include materials that are not strictly contemporaneous, and only approximate contemporaneity can usually be claimed for any two fossil individuals. A hypodigm is sufficiently homogeneous for taxonomic purposes if it displays approximately the random variation of the normal curve of frequency distribution. At least approximate contemporaneity of the individuals forming the hypodigm lends additional valuable evidence that the sample is not contaminated with homeomorphs or examples of parallelism. In most cases, an effort should be made to limit hypodigms to narrow stratigraphic units, or to single bedding planes in order to recognize the smallest stages in the evolution of a group. This may not be adequate or feasible in all cases. In the more slowly evolving animal stocks, species commonly are reported to range through an entire series or system. Careful work on such forms may show that collections from several horizons may profitably be joined into a single hypodigm which better reveals the group characters than do collections from a single horizon.

QUALIFICATIONS OF HYPODIGM SPECIMENS

One characteristic of paleontological materials in general is that they tend to be fragmentary or relatively rare. There are many exceptions, of course, but commonly it is impossible to assemble a large number of individuals from *a single horizon and locality.* Hence, a majority of fossil species were founded on "mixed samples," that is hypodigms which contain specimens certainly not of a single population. Less commonly, the original hypodigm contains representatives of two or more species, as every reviser well knows. This hazard should be recognized and noted, but probably it is not in itself adequate grounds for eliminating from consideration instructive isolated specimens if the original hypodigm is very small. It should be self evident that obscure but real and persistent morphological differences between two subspecies sometimes are recognizable only by means of adequate quantitative analysis. Hence, isolated individuals in such circumstances cannot, with certainty, be ascribed to the proper subspecies.

In consequence, any type collection which is composed of specimens from several horizons and localities is likely to lead to incorrect inferences, since it may not represent any natural group. In such an original hypodigm, I believe, contrary to Simpson's conclusion, that the members of the hypodigm should not be of equal rank. For the needs of revisers they should be clearly differentiated. One of the most convenient methods, that is already in general use, is the application of special terms to the most significant categories of "type" specimens.

TYPE TERMINOLOGY

For new work and review Simpson proposes abandonment of all type designations except the following: (Simpson, *Op. cit.*, p. 424);

Hypodigm—the statistical sample, distinguished from types and discarding the idea of types as bases for species and for comparisons.

Type—the single name-bearing and anchoring specimen. In review and revision only Simpson favors retention of the following:

Syntype—one of two or more specimens to which a single name was equally attached in the original publication.

Lectotype—a syntype subsequently selected to be sole name-bearer.

Neotype—a specimen substituted as name-bearer for one not now assignable to a definite hypodigm.

These are basic categories, but they do not satisfy all needs, in my opinion. In general it seems to me that a very useful function is fulfilled by a few additional kinds of types.

In the original type lot, or hypodigm, it is desirable to stress the circumstantial evidence that indicates which specimens possibly belong to the same subspecies or population, and which ones probably do not. It would seem sound practice to segregate the name-bearer, the holotype, or type as it is simply called by Simpson, together with all of the topotypes. This fraction of the hypodigm is likely to remain as a permanent part of the hypodigm; other specimens may be removed by a subsequent reviser. I agree fully with Simpson that paratypes (unless they also are topotypes) should be accorded no more significance than any other representative of the group in evaluating the species. In my opinion paratypes do play an important rôle. The merit in applying the term paratype to supplementary specimens in the original hypodigm is that the designation carries some prestige and is some insurance that the hypodigm will not be broken up for exchange purposes. It also stresses the published documentary value of the specimens. Implicit in the hypodigm concept is the idea that a large collection will tell more about a species or subspecies than will a small collection, and that a single type specimen will reveal very little about the natural group to which it belongs. For this reason it is urgent that large collections of a species be kept intact. I would term paratypes that come from *precisely* the same horizon and locality as the holotype topoparatypes, to distinguish them from the much less useful stray paratypes. It is clearly understood that the holotype and paratypes do not reveal any more information than subsequently described specimens of the group. Subsequently described additional specimens, in my opinion, are best segregated and designated in order to show that they also are part of a hypodigm. They are the hypotypes. To leave these published specimens without special designation endangers their future preservation. The term diplotype[3] for holotypes and lectotypes of genotype species is meritorious, it appears to me. As a name-bearer this category is pre-eminent, and a distinctive term as a safeguard is of great value for curatorial purposes.

In summary, I would recommend expansion of Simpson's list as follows:

Hypodigm
Diplotype
Holotype
Topotype
Paratype
Topoparatype
Hypotype
Syntype
Lectotype
Neotype

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[3] A few writers have consistently misused genoholotype for Knight's "diplotype." Genoholotype is the *type species*, it cannot apply to a single specimen.

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REVIEWS

Outlines of Physical Chemistry; by FARRINGTON DANIELS. Pp. viii, 713, 164 figs. New York, 1948 (John Wiley and Sons, $5.00). This book is essentially a new edition of the text formerly published under the joint authorship of Getman and Daniels. In order to provide a more logical discussion of the subject matter, the chapter on thermodynamics has been moved into an earlier section of the book where it now follows the treatment of heat capacity and thermochemistry. The material from the chapter on thermodynamic applications has been incorporated in appropriate sections throughout the text. The problems have been revised and their number increased.

As was the case for earlier editions, the aim of this book is to provide material for a first-year course in physical chemistry. Fortunately it is not directed at any particular class of students. In view of the broad coverage, the discussion of some topics (notably imperfect gases, specific heat, and entropy) is necessarily restricted. However, the clearly written, authoritative treatment of solutions and their electrical properties, equilibrium phenomena, chemical kinetics, photochemistry, and colloids, furnishes once again an excellent basis for the conduct of an introductory course. Considering the imperative limitation on length, these sections are surprisingly comprehensive.

In this new edition some of the newer topics in physical chemistry, such as radiochemistry and chromatography are treated in somewhat greater detail than before, as are the older fields of spectroscopy and photochemistry. In view of the tremendous commercial and scientific interest in polymers, the consideration of these materials is unexpectedly brief.

Professor Daniels' emphasis on the practical is indicated by the inclusion of a wide variety of problems. More than one hundred solved, illustrative examples have been inserted to supplement the text. Unfortunately, only a few of these solved problems are of the "thought provoking" type with which the exercise sets are liberally furnished.

It has been the author's expressed purpose to "raise the level of the presentation to keep pace with the improved preparation with which students are starting the study of physical chemistry." The relegation of such topics as the determination of activity coefficients to the appendix, which had already included such things as the discussion of partial differentiation, would indicate that perhaps this preparation has been leveling off.

Typographical errors, such as the omission of superscripts in the section on chemical equilibria, are infrequent and trivial. In appearance and content, this is a better book than the older editions, which have been so justly popular with student and instructor alike.

P. A. LYONS

Précis des Découvertes Somiologiques ou Zoologiques et Botaniques; by C. S. Rafinesque. Pp. 55. Palermo, Sicily, 1814 (Lithoprint reproduction, New York, 1948, Peter Smith, 321 Fifth Avenue, N. Y., $4.00).—The original edition of this early work by Rafinesque is one of the rarities of biological literature. Its reprinting makes it available widely to the many plant and animal taxonomists who are forced to consider it in their inescapable nomenclatural labors and the small printing of 350 copies should be exhausted as soon as its announcement is circulated widely.

In a prior work Rafinesque had outlined his plan for a new science "Somiologie, qui signifie la Science des Corps vivans, pour désigner collectivement la Botanique et la Zoologie unies entre'elles et régies par des loix pareilles." He designated the plant and animal kingdoms as the "Empire somiologique," in contrast to the "Empire minéralogique." Plants and animals are each divided into ten classes. The prefatory remarks provide amusing reading, but the importance of this little volume is in the systematic portion in which are described 50 new genera and 190 new species of animals and plants. Most of the material is from Sicily, but 6 new genera of plants and 45 new species of fishes and plants are from eastern North America. The names of scores of other new species are mentioned, although their descriptions are deferred until later works. As Professor E. D. Merrill states in the foreword to the lithoprinted copy, Rafinesque's work is characterized by "inconsequential original descriptions." However, it must be admitted that most of the descriptions in this volume are as significant as those of Linnaeus and even more so than those of many of the most famous later authors of the Eighteenth Century. The modern correlation of the names has not been attempted in many cases, since this work is surely unknown to most zoologists. The ten species of insects named will be almost impossible to correlate with known species. It seems equally certain that in most other groups these Rafinesquian names must become *nomina nuda.*

CHARLES L. REMINGTON

College Physics; by HENRY A. PERKINS. Third Ed. Pp. x, 786. New York, 1948 (Prentice-Hall Inc., $6.65). This new edition of a well-known and popular physics text will be welcomed by its many users. The alterations consist principally in the addition of new material, such as m.k.s. units, nuclear energy, cosmic radiation and radar. Several of the formal demonstrations have been simplified and made clearer.

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ISOSTASY AND THE ORIGIN OF SIAL AND SIMA AND OF PARENTAL ROCK MAGMAS*

WALTER WAHL

INTRODUCTION

ALTHOUGH much has been written by geophysicists and geologists on isostasy and isostatic adjustment of portions of the earth's crust, the geochemical consequences of isostatic adjustment have so far not been dealt with. This is rather surprising, since it is quite evident that the vertical change of position of rock masses brought about by isostatic adjustment will bring large masses of rock into new environments where temperature and pressure are entirely different from those under which they were formed and can remain stable, and where therefore changes of chemical equilibrium and even melting to a smaller or larger degree will take place. Thereby, necessarily, considerable geochemical changes are brought about, and these, in combination with the downward movements of portions of the earth's crust and with the rise of magmas from the deeper parts of the crust to the earth's surface, will bring about a vertical circulation of matter on a large scale in the outer parts of the earth's crust. It is the aim of the author to discuss in this paper the geochemical processes originating and taking place on account of isostatic adjustment and its extreme consequences as they manifest themselves in orogenic zones.

In a recent publication, Vening Meinesz (1948) has summarized present thought on the formation of geosynclines and on orogenic movement in the following short and lucid way: Vening Meinesz says (p. 194):

"I shall here adhere to the interpretation that the folding and overthrusting of the surface layers are the accompaniment of a great compression of the whole crust of the earth which causes a downward buckle of the crust in such a way that a great bulge is formed at the lower boundary of the crust. This bulge replaces the heavier subcrustal matter and thus represents a considerable deficiency of mass." — "The buckling limit being exceeded, the crust forms one or more waves of a few hundred kilometers in length and thus a geosyncline comes into existence, which leads to sedimentation. The process continues by means of a stronger deformation of the main downward wave which gradually forms the downward bulge described above. Incompetent surface layers are squeezed out and form folds and overthrusts at the surface." — "The further history of the geosyncline would, in outline, be as follows. We may assume that, after one or more periods of strong compression characterized by folding in the main belt, compression in the crust would decrease, and this would allow the affected area of the crust to detach itself, more or less completely, from neighbouring areas, and by rising to readjust, at least partially, its isostatic balance. This process would continue until the stage would be reached of a mature mountain range in isostatic equilibrium, as we know it in several existing mountain systems. Possibly the process would be accompanied by another one, that of the partial melting of the sialic root which, by the down-bulging, has been pressed below its normal level. We may assume that this fused matter would distribute itself below the crust and thus the root, so far as it did not disappear, would become broader and less deep."

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* An abstract of part of this paper was given as a lecture at the Geochemical Section of the 18th International Geological Congress in London, 1948.

We may add that during and after the reduction of the mountain range by erosion the underlying portion will be raised, and ultimately a peneplain is formed, which being in isostatic equilibrium, will consist of the old, formerly deeply submerged roots of the mountain range.

As the study of the great continental basement complexes of the world considered to be Archean has progressed, it has gradually been recognized that these complexes contain much material that is unquestionably composed of sedimentary and volcanic material originally formed at the surface of the earth, which material has later, by metamorphic processes, acquired its present gneissic character. The trend of thought of the school of Lyell, especially as later on developed by Sederholm, gradually gained more and more ground and an "actualistic origin" of the "Archean basement" has become generally admitted.

The rocks originally formed in an "actualistic" way have by becoming folded and overthrust been moved to great depths where remelting has taken place, in some cases of only the earlier melting or soluble portions, in other cases—when sufficient depth and temperature have been reached—of the greater portions or even the entire mass. Where we now find "the Archean" exposed it always shows signs of having been subjected to orogenic forces, and we now describe parts of it as having been formed during such or such period of mountain folding, as for instance when we in Fennoscandia speak of the rocks of the "Svecofennian Archean mountain chain." Other parts of the Archean basement here we consider to have been formed during other periods of orogenesis, in part older, in part younger than the Svecofennian movement.

It becomes more and more evident that what we are able to study of the Archean basement to a great extent belongs to what we may classify as roots of old mountain chains, the material of which has been subject to at least partial remelting or resolution or a combination of both these processes. The rock material now contained in these basement complexes has at one time, or at several times, been forced down to great depths in the earth's crust and subsequently by isostatic adjustment brought back to its present position at the surface of the earth's crust, taking part during these movements in the vertical circulation of matter in the earth's crust.

THE GEOCHEMICAL CONSEQUENCES OF ISOSTASY AND ISOSTATIC ADJUSTMENT

When the rock material either deposited by sedimentation in a geosynclinal trough or forming the basement supporting such a geosyncline is, during the development of the geosyncline and subsequent orogenic movement, forced downwards, generally to great depth, this rock material becomes exposed to temperatures and pressures entirely different from those under which it was originally formed. During movement downward of the rock material it will first undergo such kinds of metamorphosis as are consistent with the conditions reigning in the depth zones into which it enters, the well known stages of metamorphism described by Becke, Grubenmann and Niggli being passed through, and also the "facies" changes of Eskola. All the important changes that the rock material will undergo are now well known through the work of the authors mentioned above and others, and we shall not here enter upon details, but instead try to find out what happens when the rock material moves downward to depths and temperatures where the rocks begin to soften and still further down to melt.

When melting of a rock composed of several minerals sets in, it will in the case of an igneous rock start very much in a similar, though reverse, way to that in which crystallization of the rock finished when the rock was formed. That is to say, in a granite or a granodiorite the potash and soda-rich feldspar portions, and what they are able to dissolve of the quartz and the femic minerals present, will be the first parts to be liquified, and they will also absorb most of the volatile matter such as enclosed water vapor. A granitic magma will result, which at the tensions and conditions of shear existing will be squeezed out.

In a lecture given before the Geological Society of Helsingfors in 1918 the present author discussed the influence of stress on the metamorphism and partial melting of rocks (Wahl, 1918). Since only an abstract in the Swedish language has been published, a short review of the conclusions arrived at is given here:

> A fundamental difference exists between a uniform hydrostatic pressure and a unilaterally acting stress. Hydrostatic pressure raises the melting point of most solids, whereas stress lowers it, and mostly to a greater extent than it is raised by uniform pressure.
>
> In a rock which becomes exposed to strong stress, the melting points are reached much earlier for some of its minerals than for others, and especially much earlier for the later crystallized portions of the mixed crystals than for those crystallized earlier. To take an example, the portion rich in albite of a plagioclase will under the combined action of heat and stress melt earlier than the more anorthite-rich portions, which may remain unmelted.
>
> Since a rock during the action of stress is to begin with foliated at right angles to the action of the pressure, the albite-rich portions of plagioclases together with potash feldspar and quartz which they are able to dissolve will, as soon as their limit of melting is reached, be squeezed out along the foliation planes. The resulting product is then a banded rock, from which more or less of the most easily molten portions of the rock, depending on the amount of stress and on how soon the stress has been released, to a greater or lesser extent have been squeezed out. Also depending on how

soon stress is released and "regelation" sets in, the molten portions will either crystallize more or less "in situ" between the foliation planes of the rocks from which they were produced ("venites" of Holmquist) or be squeezed entirely out of the mother rock and enter the foliation planes of surrounding colder foliated rocks, forming "lit-par-lit" injections ("arterites" of Sederholm). In the latter case the gneisses remaining as residues after the squeezing out of the early melting portions will be of a much more femic character (hornblende-mica gneisses) than the original rock from which they were derived by this kind of stress-metamorphism.

Considering the general conditions in the deeper parts of the earth's crust, we may assume that, depending on conditions of depth, temperature, and stress, all gradations between "stress metamorphism" and "palingenesis" by heat action alone may occur.

A sedimentary rock or a gneissose rock containing alkali feldspars will behave very much in the same way, the alkali components, the feldspars, melting first and dissolving a certain amount of quartz and a little of femic components, forming what has often been called a "granitic eutectic." Quite generally we may say, that in any kind of rock containing alkali feldspars these will, at a certain depth in the earth's crust, start melting and a granite magma will begin to form, which, on account of its density being less than the still solid remainder of the rock, will be forced upwards.

The "palingenic" magmas originating in this way will, however, not possess much eruptive power, since they have not acquired temperatures much above their temperatures of solidification. They will therefore not be able to rise high in the earth's crust, but will probably only be able to impregnate the rock at some distance above the level from which they originate.

When the rock residues, after expulsion of the granite components, are moved downwards by further isostatic adjustments to still greater depths, there may be partial melting of the oligoclase-rich portions of the plagioclases and the dissolving of certain amounts of quartz, if such is still available in the rock residues, and also of part of the femic constituents, especially of the components rich in iron. The magma forming in this zone will be comparatively rich in femic components though a considerably "dryer" magma than those forming in a zone higher up, since these have already carried with them most of the volatiles available. Perhaps the intermediate magmas of the "charnockite series" are formed in this way, and later, while ascending, are differentiated in the series charnockite, hypersthene syenite, and hypersthene diorite.

The remaining rock mass consists now chiefly of plagioclase, pyroxene, hornblende, mica, and magnetite, forming hornblende and mica gneisses and hornblenditic gneisses. These rocks will probably, for some distance in depth, be stable and remain as such. A proof hereof is that in the granites of the basement complexes of Fennoscandia, associated with migmatites, we often find xenoliths of hornblende gneisses in which the hornblende and mica show fairly little effect of alteration or melting. At a certain depth and temperature, however, break-down of the hornblende and mica molecules will take place and a considerable quantity of volatiles will anew be freed in this temperature zone. The material, being partly molten in this zone, will principally consist of the albite portion of the plagioclase together with iron and magnesium-silicate compounds from the broken down hornblende, giving rise to the sodium-, magnesium- and iron-rich magmas containing much volatiles that we know as the "spilitic magmas."

The remaining material of the original rock ground will then at still greater depths adjust itself to a "facies" that is stable. Ultimately we may have eclogites (Fermor, Eskola), and still farther downward, when these melt, the corresponding magmas of basaltic—peridotitic character, forming the lower part of the sima.

It may here briefly be pointed out that for the sake of simplicity it has in the preceding discussion been assumed that all the potash and most of the soda have been expelled above the temperature surface limiting the depth in which granites can exist. A certain amount of both potash and soda will, however, of course be held in solution by the femic components of the residues, and the sima will thus retain a certain small amount of the alkalies.

The further question arises how far downward in the sima shell will this partial melting of the material and the rising of the molten portion upwards in the earth's crust go on? Will it continue as long as there remains any simatic crystallized silicate shell or will it cease earlier? Eskola (1936) has pointed out that when the peridotitic part of the sima shell partially melts the portion rich in iron silicate will become liquid and that this, being heavier than the magnesia-rich portion that remains crystalline, therefore ought to sink. However, conditions are actually more complicated, as we shall see presently.

When melting begins in a mixed crystal of forsterite (Mg_2SiO_4)—fayalite (Fe_2SiO_4), fayalite-rich melt will be produced. This is denser than the residue when this also is molten, but not necessarily denser than the remaining crystalline residue, which also is still rich in fayalite silicate. In the earlier part of the partial melting of a peridotite, an iron-rich silicate melt thus will be produced that will rise in the earth's crust, but gradually the remaining material will become less dense and at a certain composition the resulting molten portion and the remaining crystalline portion of the mixed crystals of the peridotite will have the same density and the molten portion will no longer rise in its surrounding of unmelted mixed crystals. When this composition of the mixed crystal is reached, crystal and melt will remain in the same layer, the liquid no more rising. A condition for equilibrium is, however, that at the same density of mixed crystal and of melt, also chemical equilibrium between mixed crystal and melt should exist. This is probably the case, since the olivine rocks enclosed in the mountain chains and the olivine inclusions in the basalts all seem to have the same composition, all containing about 10% to 12% FeO. From that temperature shell on downward where an olivine containing 10% FeO is in equilibrium with the melt, i.e., from 1,800°, the composition of the sima probably will not change further.

We arrive thus at the conclusion that isostatic adjustment and orogenesis will cause a geochemical alteration of the material of the outermost portion of the earth, causing a splitting into sial and sima layers from which volcanic material is again returned to the earth's surface. During the vertical circulation of matter thus taking place the chemical components of the rocks move in an entirely different way in the lower part of the cycle from that in which they move in the upper part. In the upper part we have the separation into clays, quartz, and the formation of carbonates; in the lower portions the rock material is split and segregated into sial and sima.

THE FORMATION OF MAGMAS DURING THE DEVELOPMENT OF GEOSYNCLINES AND OROGENIC MOVEMENTS

In the preceding chapter the general geochemical consequences of isostatic adjustment in the earth's crust have been discussed. In the following the especially extreme cases during orogenesis will be dealt with.

When sedimentation commences in a developing geosyncline, the sediments are deposited on a floor of older rock, which may consist of older sediments or of a peneplained old basement complex. If the rock floor consists of older sediments these will, however, ultimately rest on an old basement complex of metamorphic crystalline schists, gneisses, and granites. If we assume that this complex of sediments and underlying crystalline schists and gneisses, which form the supporting basement of the geosynclinal sediments, has in earlier periods taken part in isostatic adjustments, it will as we have seen in the preceding discussion, have acquired a layered constitution with a sial shell on top and sima shell farther down. The upper parts of the basement of the geosyncline will therefore consist of rocks of all kinds, but those farther down will have become layered into sial- and sima-shells and their subshells.

Now it seems to be a general rule that the first volcanic rocks to appear at the surface of the sediments of a geosyncline are rocks of the "spilitic" series, pillow lavas and picrospilitic lavas. They penetrate to the earth's surface by feeder dikes, and it seems that they appear as soon as cracks in the supporting basement of the geosyncline begin to be formed. These cracks seem to reach down to the upper shell of the sima, the temperature shell where the breaking down of the hornblende silicates of the hornblende gneisses is taking place as described earlier. Of the rock masses underlying the geosyncline this zone seems to be the first which during isostatic adjustment reaches such depths and temperature zones that geochemical changes on a large scale take place. Its material breaks down partially and forms a magma rich in volatiles, which is of great liquidity and mobility and therefore easily ascends through the cracks to the surface as lavas. It appears from our present knowledge of the volcanics erupted during the early phases of a geosyncline that these liquid spilitic magmas containing much volatiles dominate strongly at certain places; however, at some places they seem to be accompanied by smaller amounts of rhyolitic rocks which probably originate from the temperature shell where the melting of the alkali feldspars sets in. It is characteristic of some geosynclines also that basaltic rocks alternate with the spilitic, the basaltic magma probably rising through cracks that penetrated farther down through the upper sima and reached the basaltic subshell of the sima.

If we study the development of magmatic action during the formation of a geosyncline we find, however, that at a certain stage the spilitic volcanism entirely ceases and a period of intrusive igneous action sets in, resulting in the formation of large laccolithic masses of granodioritic magmas which are injected concordantly into the sediments deposited in the geosyncline.

Now how is this change to be explained? As Vening Meinesz says in his summary of the development of a geosyncline, quoted above, after a given period the horizontal compression in the crust will exceed the "buckling limit," the process resulting in a downward bulge of the rock masses forming the supporting basement of the geosynclinal sedimentation. This down-bulging brings great rock masses down into temperature zones that are at a much higher temperature than those under which the rocks were formed; probably far down into the sima shell where temperatures are sufficiently high to melt the sial rocks entirely. When large portions of all the rocks of the basement complex are thus melted down, magma masses in great quantities will be formed, which will have a composition corresponding to the average composition of the basement complex. Now Sederholm (1925) in his well-known paper on the average composition of the earth's crust in Finland, has calculated that the composition of the Archean terrane is similar to that of the average granodiorite as calculated by Daly. The granodioritic magmas, which result from the entire basement being driven down deep into the sima shell and becoming molten, will possess the high temperatures of the sima shell and contain fairly large amounts of volatiles. On account of this condition they will also possess a high amount of eruptivity, and since they will be available in large quantities they will form the principal magma of the earlier part of an orogenic period ("primorogenic granites," Wahl, 1936), as we for instance find them in the great central batholiths of Alaska, through British Columbia and the Coast Ranges, further through the Andes of South America and down as far as Tierra del Fuego, and also as the andesite lava of the whole chain of volcanoes along this and other orogenic zones.

The production, mobilization, and intrusion of the granodioritic magmas will continue as long as the bulging down of the geosyncline takes place. But when isostatic adjustment again sets in, and the mountain range is rising, intrusion of granodioritic magmas ceases, although the ascent of andesitic magmas at the summits of the mountain range and the activity of the andesite volcanoes continues for some time longer.

During the subsequent course of the orogenesis stretching will occur and cracks will be formed, reaching far down, permitting basaltic magmas of simatic character and origin to rise. This period corresponds to the formation of the "basic dikes" (Sederholm) that are so well exhibited in the old Precambrian mountain chains.

Later, during further orogenesis, strong lateral compression again manifests itself, resulting in folding and overthrusting, whereby large rock masses are piled upon each other and deeper-lying portions are again forced downward. During this stage of orogenesis, which often gives rise to overthrusting and movement on an Alpine scale and style, the granodiorite plutons, which during an earlier stage were concordantly intruded into the geosynclinal sediments, as described above, are in their turn folded together with the surrounding sediments and afterwards appear as "older granites" (Sederholm) of the terrane. When this rock material during the height of orogenic movement in its turn is forced downward, it will, together with other rock material, reach depths where remelting and resolution of the early melting components of the rocks takes place and large quantities of potash-rich granite and pegmatite magmas are formed ("serorogenic granites," Wahl, 1936). This zone is also the zone of formation of the migmatites. Some of the granitic matter will probably not be squeezed out entirely from the original rock mass and the process may result in the formation of more or less schistose gneisses with feldspar-quartz layers between layers of the dark minerals, or the formation of "venites." The juices of granitic composition will in other cases be squeezed out of the rock mass and on account of their small density will rise to a higher level in the orogen. The greater portion of these granitic magmas will, however, probably not travel far, but as soon as they reach rock masses no longer soft, but cracked and schistose, will be sucked up between the layers of schistose or strongly pressed rocks by capillary force or by being pressed in, "lit par lit," between the foliation planes of the schistose rocks, forming "injection gneisses."

In those cases where the whole mass by repeated isostatic adjustments in both directions moves up and down, the zone of migmatite formation in the root portion of the mountain will also move up and down, the same rock material repeatedly passing the active temperature zone. The zone where migmatite has been formed thus may pass up and down, and the uppermost limit that it reached in the rock ground may then give the impression of a "migmatite front." When the migmatitic rocks move downwards on account of isostatic adjustment or by being bulged down they will not as such remain unaltered beneath the temperature shell in which they were formed, the salic material of this temperature surface, when the rock masses pass downward, always being melted anew and rising higher.

The granitic and pegmatitic material originating from the migmatite zone will, when penetrating upward, cut both the spilites, the granodiorites (Archean "older granites" of Sederholm), and the "basic dikes," thus appearing as younger granites in the orogenic zones (Archean "younger granites" of Sederholm).

During orogenesis there thus appear four distinct magma types that in a sense can be regarded as "parental magmas" from which by subsequent differentiation other rock types branch off. We have, firstly, the spilitic rocks grading over into picrites; secondly, granodioritic magmas that differentiate to magmas covering almost the whole range from granites to peridotites; thirdly, basalt magma; and lastly, potash granitic magmas of a composition very close to the granite eutectic. These rock magmas represent the result of the downward movement of the rock-material in the earth's crust and the subsequent rise to the surface of the earth of magmas formed in depth, the whole forming a cycle of vertical circulation of matter in the earth's crust. When the rising magmas have solidified as rocks, these again become subject to erosion and form the material for sediments in new geosynclines. The rocks of these will again be subject to orogenesis and the material again in the depths split into the four magma groups mentioned above, the vertical circulation going on and being completed once in every orogenesis in which the material takes part.

THE MAGMAS OF THE "CRATOGENS"

Orogenesis in its later phases often passes over into crustal movements of a more "cratogenic" character and such movements may also occur on either side of the orogenic belts. The igneous activity in the cratogenic zones is of a different type from that of the orogenic zones previously discussed. The magmas appearing in the cratogens consist principally of plateau basalts, of the basalts rich in olivine of the oceanic hemispheres, of the alkalic foyaitic and essexitic magmas, and of a group of mostly potash-rich granites and with them associated gabbros, norites, and plagioclase-rich rocks (the "Post-archean" igneous rocks in Fennoscandia, the granites, red rocks and associated gabbros of the Keweenawan, the rocks of the Scotch igneous Tertiary provinces, etc.).

The plateau basalts, which apparently are the rocks of this group which appear in the largest masses, originate probably from the upper zone of the lower part of the sima, whence they rise when under cratogenic conditions cracks open up to great depths.

The olivine-rich basalt clan probably takes its origin from a still deeper subshell of the sima.

The granite-gabbro associates of the Tertiary provinces of Great Britain and some of the "Post-archean" igneous rocks in Fennoscandia probably originate by differentiation from parent magmas of composition similar to the granodiorites but of a much "dryer" character. It seems as if the intermediate differentiation products, diorites and mica diorites, etc., are left out during differentiation of such magmas "dryer" than the granodioritic magmas proper, the associations granite-gabbro and granite-anorthosite being formed.

Such a development of a dry magma in which only subordinate amounts of mica and hornblende are formed can theoretically be explained by diagrams of a type similar to Bowen's diagram (Bowen, 1928, Fig. 59). A granodioritic material from which some part of the granitic constituents has been separated by squeezing out, but which still contains orthoclase and mica would on further partial melting in a temperature zone where mica breaks down yield a magma that on differentiation splits into potash granite ("rapakivitic magma") and norite-anorthosite.

Possibly also some of the "charnockites" are more closely connected with "cratogenic" conditions than with orogenic.

There remains further the question of the origin of the alkalic rocks. As recognized long ago by Becke, the alkalic rocks are associated with the cratogens. The alkalic magmas are comparatively dry magmas, but in spite of this they show great fluidity and ability to differentiate. They therefore probably originate in deep-seated, comparatively hot regions. Several attempts have been made to explain the formation of such undersilicated alkalic magmas by chemical reaction of subalkalic magma with desilicating agents, the hypothesis of Daly that desilication has taken place on account of reaction with limestone being the most important one. It has, however, been shown that serious objections can be raised against this explanation.

It seems that a cause of the formation of alkalic, undersilicated rocks could be that the original material has been rich in mica and hornblende, the component of the micas corresponding to kaliophilite silicate acting as desilicating agent. Mica, which is stable at fairly high temperatures, when sufficient amount of water vapor is present, breaks down on dehydration and melting and splits into olivine and kaliophilite silicates, of which the latter at high temperatures reacts with albite and desilicates part of the albite to nepheline, while itself changes to leucite and orthoclase. The desilicating action of molten biotite and the formation of some components of alkalic rocks is demonstrated by the experimental work of Fouqué and Michel-Lévy, who on melting about equal amounts of mica and microcline obtained a product consisting principally of crystallized leucite and olivine, together with some melilite and glass. The kaliophilite silicate on reacting with the orthoclase silicate, changes to leucite and desilicates the orthoclase to leucite. Also hornblende on breaking down delivers material of alkalic magmatic composition; the "norms" of most hornblendes from subalkalic rocks containing considerable amounts of nepheline and leucite, besides the chief normative constituents anorthite, olivine, diopside, and albite. Simultaneously with the partial desilication of albite, chemical equilibria among the femic silicates are also readjusted, the resulting pyroxenes and amphiboles taking up large quantities of sodium ferric-iron silicates instead of the lime-alumosilicates contained in the subalkalic rocks.

Since biotite and hornblende are evenly distributed all through the bulk of the original rock material, their components will be present everywhere in the resulting molten magma and can therefore directly act as desilicating agents. Consequently, there is in this case no need to assume that the desilicating agent, as in the case of limestone, has been assimilated and by diffusion has been able to penetrate the whole magma mass, thereby permitting a uniform desilication and causing the formation of an alkalic magma.

Bowen has given approximate diagrams explaining the formation of nepheline from a subalkalic magma by the intermediate formation of leucite (Bowen, 1928, Fig. 64). An analogous diagram, in which we substitute the kaliophilite silicate (from broken-down mica in a dry magma) in place of the leucite silicate, would still more efficiently explain a magmatic development leading to foyaitic magmas. In a magma containing a sufficient amount of volatiles the equilibrium:

$$NaAlSi_3O_8 + KAlSiO_4 \rightleftharpoons NaAlSiO_4 + KAlSi_3O_8$$

will shift towards the left as more of the kaliophilite silicate is used up in forming mica, whereas in a dry magma, where the amount of kaliophilite silicate is augmented by the breaking down of mica, the equilibrium will shift to the right, more nephelite being formed. It is, however, not possible here to discuss further this special problem.

Also, when masses of diorites, especially mica diorite, or of andesites, which are initially free from quartz, are depressed to such depths that partial melting sets in and dehydration also has proceeded far, the melt will consist chiefly of the albite component of the plagioclases in which are dissolved the components from the broken-down micas and amphiboles, and the conditions requisite for shifting the chemical equilibrium from the state dominant in the subalkalic magmas to that dominating in the alkalic magmas will thus exist in these molten portions. By the squeezing out of such molten parts alkalic undersilicated magmas of the nepheline-phonolite clan may originate.

Just as diorites and andesites in the way described above may yield nepheline syenite magmas, gabbros and basalts would in a similar way yield certain amounts of essexitic and theralitic magmas. The residue from such squeezing out of the sodium-rich portions of diorites, if melted at greater depths, give rise to the pyroxenitic magmas that often accompany nepheline syenites and essexites. It is interesting to note that from a theoretical point of view every kind of rock material containing much biotite, or biotite and hornblende, by being depressed to a zone where the temperature is high enough to cause the melting of the rock and the desilicating reactions to take place, ought to change to phonolite magma. The greater the portion of biotite of the original magma, the richer in leucite will be the alkalic rocks thus formed.

THE ORIGIN OF PARENTAL MAGMAS

There is a strong tendency in modern petrology to attribute the formation of the different igneous rocks to a common source, a parental magma, from which by differentiation all igneous rock types are thought to have been developed. The plateau basalt magma is mostly regarded as the parental magma of all other magmas, and it has been shown by Bowen (1928) that it is possible theoretically to derive the principal types of magmas by crystallization - differentiation from basalt magma. Several objections have, however, been raised to the practical application of this principle to all petrogenetic problems. Also peridotite has been considered by some authors to be the parental magma of basalt and thus of all other magmas, and Kennedy (1933) has attempted to simplify theoretical deductions by postulating the existence of two different basaltic parental magmas, viz., a tholeiitic and an olivine basaltic.

The processes of vertical circulation of matter in the earth's crust here discussed lead to the conclusion that layers of different composition are formed in the outer crustal portion of the earth, which layers may be grouped in two principal shells, the sial and sima shells. When sublayers of the sima or sial move to deeper zones where partial melting takes place accompanied by squeezing out of magma, we have seen how magmas of different kinds are formed, and under different phases of both orogenic and cratogenic development are forced to the surface of the earth. Owing to the conditions of their formation, these magmas will mostly possess a sufficient amount of heat and liquidity to differentiate during their rise through the outer part of the earth's crust. We thus arrive at the conclusion that as a result of the chemical consequences of isostatic adjustment several principal types of "parental magmas" are generated, each in a particular sublayer of the sial, salsima, or sima, and that the generation of one parental magma, as for instance the granodioritic, is independent of the generation of another, such for instance as the basaltic. The conditions of evolution of a geosyncline, an orogen, or a cratogen are the cause of the formation and rise toward the earth's surface of each particular parental magma.

During geosynclinal development and orogenesis at least the four following different kinds of parental magmas are formed: 1, spilitic and picrospilitic magmas; 2, granodioritic and dacitic-andesitic magmas; 3, basaltic magma; and 4, granitic magma. Under cratogenic conditions we obtain further 3, basaltic magmas of which the "tholeiitic" probably originate from the upper zones of the sima and the "olivine basaltic" from somewhat greater depths; 4, granitic magma, differentiating into potash granite and gabbro-norite-anorthosite; 5, charnockitic magmas, differentiating into hypersthene granite, hypersthene syenite and hypersthene diorite and norite; and 6, foyaitic magmas, by differentiation giving a wide range of alkalic magmas. From the lower part of the sima shell, 7, peridotitic and pyroxenitic magmas originate. All these magmas may be considered as parental magmas, which by further differentiation give rise to the large number of igneous rocks met with.

During this discussion it also becomes clear of what fundamental importance for the further development of the different magmas the previous dehydration and the degree of "dryness" of the parental magma is. It cannot be too much emphasized how fundamentally important for the trend of magmatic development the formation of the mica and hornblende silicates is. Until the important research work on silicates containing volatiles, now being conducted at the Geophysical Laboratory of the Carnegie Institution, has proceeded further there seems to be little possibility of giving an adequate and quantitative description of the evolution of the magmas crystallized under abyssal conditions.

THE ORIGIN OF METASOMATIC FLUIDS IN THE BASEMENT COMPLEX

Earlier it was pointed out that in a certain zone of the upper sima a large amount of rock residue rich in hornblende and mica will occur and after partial melting will furnish the material to form the spilitic magma. But when rock masses containing hornblende and mica, by isostatic adjustment or by bulging down, approach or reach this temperature zone certain amounts of the volatiles will necessarily escape even before the actual melting of the silicates. Also portions of the volatiles in the supercritical state may separate as supercritical fluids not soluble in the silicate magma portions and will travel upwards in the basement complex of rocks. All these fluids will be able to cause intense chemical changes of "pneumatolytic" character.

Recent investigations on the sedimentation in high pressure turbines, an engineering problem of considerable importance, have shown that water vapor at high temperatures in the supercritical state takes up and can hold surprisingly large quantities of iron-sodium-silicates (acmite) and also of sodium and aluminum silicate (Burkhardt and Imhoff, 1947; Morey, 1942). The occurrence in teschenites of chlorite crystallizing later than analcite (Campbell, Day, and Stenhouse, 1932), clearly shows that water vapor at high temperature dissolves also large quantities of magnesium and iron silicates.

Isostatic and orogenetic movements will thus produce also large amounts of the volatile agents responsible for the metasomatic processes in the earth's outer crust.

THE ORIGIN OF SIAL AND SIMA

We have seen above that isostatic adjustment and orogenic downwarping of the rocks of a geosyncline and its supporting basement rocks will result in the partial melting at certain depths of the easiest melting portions of the rock foundation, the feldspars. This molten feldspar will dissolve certain amounts of water vapor and also of quartz and of a small amount of the femic constituents of the rocks. Magmas of granitic composition will thus be formed. These magmas will be much less dense than the remainder of the rock foundation and will therefore rise and travel upward. Granitic rocks and magmas, therefore, can exist in the outer crust of the earth only above a certain temperature surface and it is physically and chemically quite impossible that granite magmas should invade the outer parts of the earth's crust from deep down in the earth as generally has been the view earlier held by geologists. That the granites when brought to a sufficient depth in the earth's crust will entirely melt and be regenerated as magmas has especially been advocated by Sederholm (anatexis, palingenesis) and views related to those expressed in the present paper have earlier been published by Wahl (1918, 1936, 1943), von Wolff (1928), Eskola (1932, 1936), Holmes (1932), Kennedy and Anderson (1938), Stille (1939). Lately also Daly (1943) has recognized that granitic magmas cannot exist in the earth's crust below a certain depth. The ideas expressed in the papers mentioned above have, however, not been connected with the problems arising out of isostatic adjustment and the resulting vertical circulation of matter, and I think it is evident from the preceding discussion how intimate the interdependence of all these processes are, and that a full understanding of the formation of sial and sima is not possible without considering all these problems together.

The temperature surface in the earth's crust beneath which the melting of the alkali feldspars takes place represents therefore also a dividing surface in the earth's crust above which the granite material accumulates. This surface thus represents the bottom of the "sial layer." Since there is a difference of about 125° between the melting points of orthoclase and of albite in the presence of water vapor, the melting of the alkali feldspar components of the rocks will actually take place in a temperature shell of about 4000 m. thickness, the geothermal gradient here being taken as 30° per kilometer. We may approximately calculate the position of this zone from the melting points of albite and orthoclase and their mixtures. From the data in Goranson's investigations (Goranson, 1931, 1938) we arrive further at the conclusion that an albite-orthoclase-quartz mixture of granitic composition, saturated with water vapor, becomes liquid at about 800° to 925°. From these data we estimate a depth for the lower bordering surface of the sial of 27 to 31 kilometers. Under the influence of unilateral pressure and shearing the local melting of the early melting components of the rocks may, however, take place at much lower temperatures, i.e. at higher levels in the earth's crust, and the zone of formation of migmatites and granitic magma will under such conditions be extended upward to higher levels. Also the thickness of the sial shell in different parts of the earth must vary according to the variations of the geothermal gradient and be different under continental areas and orogenic belts. From earthquake data it has been assumed that the sial layer below mountain ranges continues downwards to a depth of 35-40 km. (Gutenberg, 1943, 1945), but in other parts of the earth's crust may be considerably thinner (down to 20 km.).

The portions of the rock material remaining after the granite components have been expelled, principally consisting of femic components of the rock foundation, will in zones farther down in the earth's crust have to adjust themselves to "facies" that are stable at the higher temperatures and pressures prevailing there. We may regard the "sima" shell as having been formed by these rock residues and, as we have seen above, such a sima shell must consist of different layers. The outermost of these layers still containing plagioclase of intermediate composition and minerals of the hornblende and mica groups should perhaps more correctly be regarded as a separate "salsima" shell corresponding to Gutenberg's "intermediate layers" (Gutenberg, 1943).

The sima proper, beneath the "salsima," will also be composed of subshells, the outermost being of basaltic character underlain by pyroxenitic and peridotitic subshells. With regard to the thickness of the "salsima" and the outer shells of the sima, these will vary with the geothermal gradient in different parts of the surface of the earth. The peridotitic shell, as explained earlier, will probably at a certain depth acquire a constant composition, the olivine there containing about 10% to 12% FeO. Olivine of this composition is at ordinary pressure in equilibrium with its melt at a temperature of about 1780°. This temperature, however, will be raised by pressure to about 1900° to 2100°, corresponding to a depth of 65-70 km. Gutenberg gives the depth of the Mohorovicic discontinuity as at 55-65 km. beneath mountain ranges and as low as 40-50 km. under the continents. From here downward to a considerable depth the velocity of the longitudinal wave remains constant at 8.0 km./sec. This figure corresponds to the figure for peridotite obtained experimentally, and we are thus entitled to assume that from here to great depths the sima shell has a constant chemical composition and consists of olivine and molten olivine in equilibrium, the molten part perhaps being of rigid glassy nature, in view of the high pressures prevailing at these depths. The conception arrived at here as to the constitution of the sima shell differs considerably from that proposed by Eskola (1936), but approaches in many respects the views of Gutenberg and of Daly, as set forth in Daly's latest publication (1946).

THE PROBLEM OF THE ORIGINAL CRUST OF THE EARTH AND THE LAYERING OF THE OUTER PORTIONS OF THE EARTH

Before concluding this discussion I should like to consider briefly the problems, have the sial and sima shells been developed by layering of the material before the outermost crustal portion of the earth had solidified or have these shells been formed later, and of what kind of material was the original crust of the earth formed?

The process of vertical circulation of matter in the outer part of the earth's crust, as outlined in this paper, shows that rock material of very different kinds is split into sial and sima and added to these layers. At least part of the sial and sima shells of the earth's crust have, therefore, been gradually developed out of various rock materials during the geologic history of the earth and these shells are thus not necessarily formed by layering of the original molten material of the earth.

In a paper on the Chemistry of the Meteorites the present author concluded that the chondritic meteoritic stones were formed as tuffs constituting the first "crust of solidification of a cooling cosmic body." It was pointed out that if surface conditions had been similar to those on the sun, at a later stage of development than the present, tuffs containing chondrules might be precipitated from the hot atmosphere onto a surface that was beginning to crystallize, and that deeper, beneath these surface layers, the molten material might crystallize as rocks like the various kinds of "achondrites" known (Wahl, 1910). Recently Daly has published an extensive paper on "Meteorites and an Earth-Model" (Daly, 1943).

Most petrographers who have studied the chondritic stony meteorites, which form about 80% of all stony meteorites, agree that these meteorites are of a character similar to volcanic tuffs and tuff breccias. The friable spherulitic chondrites, which are very numerous and consist of tuffitic assemblages of chondrules, mineral fragments and fragments of silicate glass, represent probably a primary stage of development, from which the more "crystalline chrondrites" and brecciated and veined chondrites have developed by various kinds of metamorphic processes. The principal changes seem to have been brought about by heat effects causing a partial sintering and an increase of crystallinity, and such alteration of friable chondritic matter into crystalline chondrites has been reproduced by prolonged heating of pieces of the chondrite of Bjurböle to temperatures not quite reaching the melting points of the constituents (Wahl, 1910).

If we assume that the original crust of cooling of a cosmic body like the earth was not a kind of gneissose granite, as often pictured in early geologic literature, but consisted of accumulations of chondrums and mineral and glass fragments constituting chondritic tuffs, grading downward into more "thermo-metamorphosed" material such as the "crystalline chondrites," and still farther down to crystalline material like the achondritic meteorites, the "chassinites," "chladnites," "rhodites," and meteoric "eucrites," which in structure resemble terrestrial ultrafemic rocks like peridotite and eucrite, can we explain the formation of sial and sima out of such meteoritic material?

The stony meteorites consist principally of magnesium and iron silicates, but according to the calculations of the average composition of all stony meteorites recently published by Brown and Patterson (1947-48) they contain 3.45% Al_2O_3, 2.90% CaO, 1.10% Na_2O, and 0.25% K_2O.* The chondritic stony meteorites thus contain so much aluminum and alkalies that conceivably these constituents, together with part of the lime, magnesia, iron oxides, and silica, enriched and separated from the main bulk of magnesium and iron silicates, could yield sufficient material to form a sial shell. If we estimate that four or five parts of a chondrite of average composition can yield one part of sialic composition, that is, of granodioritic composition, this seems to suffice for the production of the material of the sial shell.

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* Unfortunately, our present knowledge of the average composition of the stony meteorites still remains rather inadequate. More exact knowledge can be attained only after a sufficient number of trustworthy analyses of meteorites belonging to the different groups of the stony meteorites is available and a weighted average is calculated, taking into account the number of falls in each group.

The sima and sial shells could gradually have been developed from a primary crust of chondritic composition as soon as the earth had cooled below the critical temperature of water, enabling water to be condensed from the atmosphere, and erosion and decomposition of the rocks at the surface of the earth had begun. In due time compression and movements in the crust would have set in and present conditions of vertical circulation would have developed. Gradually quartz and clays would have been formed at the surface together with carbonates of magnesium, iron, and calcium. This material, together with the remaining chondritic, would then, when the formation of geosynclines began, have been moved down to sufficient depths and by partial melting, in the way described above, magmas like our present ones would have been formed. These would in part have risen to the surface, the sima material remaining below, and in this way, as soon as the present type of weathering at the earth's surface and the formation of geosynclines had set in, a change of matter from the chondritic conditions to the present ones would also gradually have set in and sial and sima been developed. If evolution in the outer parts of our planet was of this type, it seems possible that beneath the sima portions of the meteoritic type of matter may still remain. The age of the crust of our earth is already so great that there is little doubt that the change from original meteorite material in the crust to present material has had time to proceed far and yield the present quantities of sial and sima.

According to these views the outer portion of the earth has consisted of a "crust" formed of meteoritic chondritic tuffs. Only later during geologic evolution has this crust as the result of geologic processes and the vertical circulation of matter acquired its present layered constitution, and sial and sima and the rock magmas and rocks now found in the outer portions of the earth been formed.

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HELSINGFORS, FINLAND

PROCESSES OF EROSION ON STEEP SLOPES OF OAHU, HAWAII

SIDNEY E. WHITE

ABSTRACT. The small particles of weathered rock debris derived from the lava flows and associated pyroclastic rocks of the Koolau Range, Oahu, Hawaii, are transported from the mountain slopes of the Range to the main streams below by the processes of soil flow and soil creep. The particles are moved slowly and progressively down through the ravines and across the cliffs of the leeward theater-like valleys, and over talus-veneered bedrock slopes to the valley floors. The occasional mud flows that break out through the forest vegetation contribute to the soil flow.

Variegated patches of old thick soil, maturely developed soil, youthful thin soil, or bare bedrock on the 70-80 degree slopes at the head of one leeward theater-like valley are considered evidence of avalanching and are important in the recession of these steep slopes. A cycle of recurrent sliding is suggested, similar to that postulated by Freise for steep slopes in the Brazilian tropical forests. This cycle includes the development of soil on the 70-80 degree bedrock slopes with the establishment and growth of plants, eventual soil exhaustion after maximum use, instability induced by rainfall, and avalanching with exposure of the bedrock again.

INTRODUCTION

KOOLAU Range forms the eastern mountain mass of Oahu, Hawaii (fig. 1). Basaltic lavas and associated pyroclastic rocks of the range are built up in continuous succession into an elongate volcanic dome. The western side of the island is composed of an older volcanic dome, the Waianae Range. The two mountain ranges merge in a low, saddlelike plateau. Secondary cinder cones, lava cones, and ash cones along the south shore and inland constitute the more recent features of volcanic activity on the island. Further geographic description is not necessary; adequate accounts are found in Stearns and Vaksvik (12),[1] and Wentworth and Winchell (17).

Much of Koolau Range has undergone severe erosion and very little, if any, of the original surface of the volcanic dome remains. A few highland flats near the summit divide may reflect an original surface, and possibly several gently sloping, triangular facets on the southwest side of the Range may represent in a geomorphic sense the original flow slopes (13, p. 8-9; 17, p. 55). Waianae Range is much more extensively dissected throughout.

A great cliff extends along the northeast or windward side of Koolau Range from the southeast tip of the island north to Kahana Valley, a distance of approximately 25 miles. This cliff or precipice probably owes its existence to a combination of processes of erosion which include (1) extensive headward erosion in stream valleys aided by relatively greater amounts of rainfall at higher altitudes, (2) capture of ground water from the southwest side of the divide of the Koolau Range by the deeper eroding northeast side streams, (3) steepening and possible undercutting of steep slopes and cliffs by springs,

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[1] Numbers in parentheses refer to references at the end of this paper.

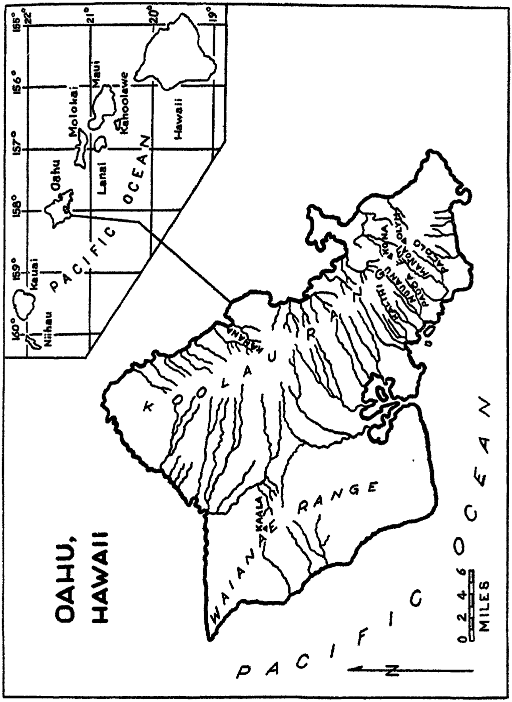

Figure 1. Index map of Hawaiian Islands, and map of Oahu, Hawaii. Kalihi Valley, Nuuanu Valley, Pauoa Valley, Manoa Valley, Palolo Valley, and Kahana Valley are indicated. Kaala—Mount Kaala, 4025 feet, is highest peak of Oahu and of Waianae Range; Kona—Puu Konahuanui; Olym—Mount Olympus.

(4) landsliding and soil avalanches, (5) coalescing of valley headwalls into one irregular, crenulated precipice, (6) removal of narrow residual spurs between valleys by marine abrasion following submergence, and (7) emergence leaving broad, flat valley floors and some valleys deeply filled with sediments (12, p. 28-29 and Pl. 8). On the southwest or leeward side of Koolau Range, however, no such great amount of erosion has occurred. There, some of the valleys were eroded so deeply, to a base level lower than the present, that their bottoms and lower sides are aggraded to form an apparent U-shaped valley. Others, cut less deeply, are correspondingly less wide and lack the aggradational features in the bottom, and appear V-shaped (16).

A study of the processes of erosion on the steep slopes at the heads of several great theater-like valleys on the leeward side of Koolau Range constituted the main purpose of the field work presented by this paper. For possible comparison by the reader with other semitropical and tropical regions, a brief discussion of the climate, vegetation, and soil of Oahu is first reviewed. Description and discussion of the origin of the theater-like valleys is included.

ACKNOWLEDGEMENTS

The writer wishes to express his indebtedness to Dr. Kirk Bryan of Harvard University who first called his attention to the general lack of precise knowledge concerning slopes in tropical regions during a series of lectures in 1940. Dr. Chester K. Wentworth of the Honolulu Board of Water Supply has kindly reviewed this paper, and Mr. Charles E. Stearns of Tufts College has read the manuscript and made various suggestions for its improvement. The original field work was conducted while the writer was stationed on Oahu in 1944; discovery at a later date (and subsequent study) of the publication *Soil Avalanches on Oahu, Hawaii*, by Wentworth, a paper which largely anticipated the writer's field work, has stimulated this publication.

CLIMATE

The Hawaiian climate, generally speaking, is dominated by two elements: trade winds and orographic rainfall. Winds blow more than 80 per cent of the time from the northeast and east. Uniformity of temperature is especially noticeable. Altitude continually controls both temperature and precipitation.

The North Pacific subtropical anticyclone is the immediate source of the trade winds. These oceanic winds, however, are relatively stable when they reach the Hawaiian Islands, are of shallow depth and give little rain; the effects of orographic lifting are the cause of precipitation. The majority of rain showers are heavy, most intense over mountains, some quickly dying away to leeward, others general over both islands and ocean. Cyclonic fronts, however, produce the most changeable weather, known by the Hawaiian word *kona* for leeward or southwest, and the islands' winter rainfall maximum. Kona weather is characterized by rain-producing fronts of short duration associated with high humidity and southerly winds, often followed by heavy precipitation; coastal regions on the southwest sides of the islands are most affected (9, pp. 29-51).

Figure 2 shows the distribution of rainfall on Oahu, taken from Carson (2). Average annual rainfall on Koolau Range varies from less than 30 inches to more than 300 inches. Part of Waianae Range lies in a rainfall shadow caused by Koolau Range. The amount of rainfall on Koolau Range increases with altitude, and with distance west or southwest of the Range crest. As a consequence, maximum torrential precipitation falls not upon summit crests but to leeward, as shown by figure 2. Annual precipitation for certain leeward valleys along the southwest side of Koolau Range, such as Kalihi, Nuuanu, Pauoa, Manoa, and Palolo Valleys, varies from 40 to 180 inches. Rainfall along the length of any one valley also varies. For Manoa Valley, approximate annual average at the valley portal is 41 inches, in mid valley 90 inches, on the upper valley floor 176 inches, and near summit ridges of the Range 106 inches (10, p. 566). With such precipitation, main streams never run dry. Water enters the main valleys from steep tributaries by means of waterfalls and cascades, and flows from series of springs along the foot of headwalls and cliffs. The waterfalls and cascades, however, are usually ephemeral, flowing only a few hours or a few days after a rain.

VEGETATION

Rainfall is the controlling factor in the phytogeography of the leeward valleys and of the Hawaiian Islands in general.

Manoa Valley, for example, becomes progressively hygrophytic headwards (10, p. 566). MacCaughey (10, p. 569) has divided Manoa Valley into ecologic zones: (1) Valley Floor

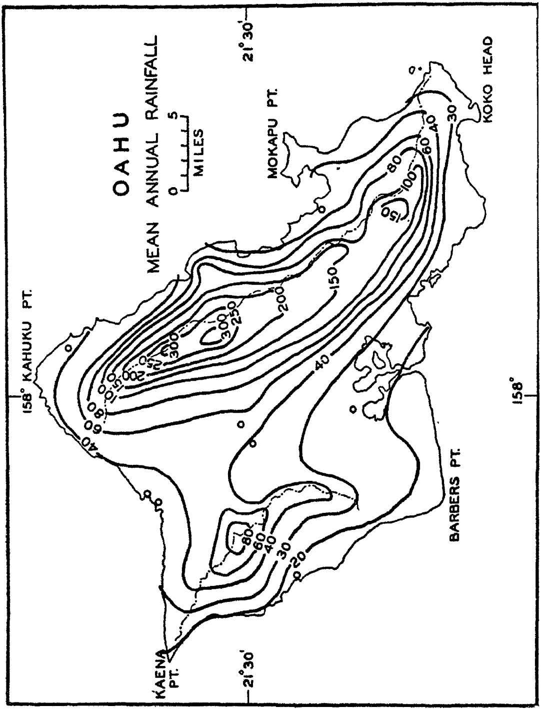

Figure 2. Distribution of mean annual ra infall on Oahu, Hawaii. Taken from Carson Dot-dash lines represent the crests of Waianae Range and Koolau Range.

(lower floor near portal, upper floor near head), 50-300 feet in altitude; (2) Manoa Stream and its tributaries; (3) Talus zone, 100-300 feet; (4) Valley Walls or Lateral Ridges, 50-1000 feet and 1000-2000 feet; (5) Kukui zone (ravines and precipices), 300-1000 feet; (6) Koa and Lehua zone, 1000-1400 feet; (7) Hanging Valleys (rain forest), 1000-3000 feet; and (8) Summit Ridges and Peaks (including Mount Olympus, 2330 feet, and Puu Konahuanui, 3150 feet), above 2000 feet. Other leeward valleys may be divided into similar ecologic zones.

Valley floors near the heads of the valleys are heavily forested with a jungle of hau trees (*Hibiscus tiliaceus*) especially adjacent to the stream channels. Groves of the guava tree (*Psidium guajava*) are found on higher slopes, along with *Lantana camara*, *Eugenia malaccensis*, *Cordyline terminalis*, *Verbena litoralis*, and *Pandanus odoratissimus*. The steeper, upper valley walls and the walls of ravines are covered with luxuriant kukui groves (Kukui or Candle-Nut tree, *Aleurites moluccana* Willd.) which often form light green bands across the heads of the valleys. Steep slopes and precipices at the heads of the theater-like valleys support heavy shrubby growth, but may also be covered with hilo grass (*Paspalum conjugatum*). Above the kukui groves is the zone dominated by koa trees (*Acacia koa*), lehua trees (*Metrosideros collina*) and other rain forest types. Undergrowth is scanty at this altitude, consisting mainly of bryophytes and lesser species of pteridophytes (10, pp. 593-596). A few grass forms exist.

SOIL

In general, on lower Koolau Range slopes, a sublateritic soil exists in contrast with that on the upper slopes. Oxidation and relative accumulation of iron, with little soil movement, is indicated on lower slopes, whereas, with lower temperatures and greater rainfall, reducing conditions and the loss of iron are possible on upper slopes. In Hawaii, the soil absorbs rainfall and is relatively resistant to soil erosion. With a heavy forest cover, a thick soil mantle is maintained, which can stand even on very steep slopes. In the high, upland areas of Koolau Range, however, soil materials are moved so rapidly that normal soil development is rare (17, pp. 63-64).

In the lower valley bottoms, thick deposits of old alluvium are thoroughly weathered, stable, somewhat indurated, and rammed tight by weathering expansion. On high mountain slopes the alluvium is less weathered, much more permeable, and is much younger than that in the valley bottoms. Soil and mantle rock are more susceptible to slow creep and soil avalanching in the mountain areas (17, p 65; 16). The chief soil masses on the mountain slopes in the heavy rainfall belt are only 10 to 30 feet thick (16); on the steep slopes of 70-80 degrees the soil is considerably thinner.

THEATER-LIKE VALLEYS

Mature looking, U-shaped theater-like valleys, eroded by consequent streams into leeward slopes of Koolau Range, are rimmed by steeply pitching ravines and precipitous walls (fig. 3). These valleys lack important tributaries. The floors of many of the valleys near the portals are flat surfaced and filled with hundreds of feet of alluvium. This partly accounts for their mature appearance. Stearns (11, p. 15) states that "Manoa Valley is a typical amphitheater-headed valley carved by stream erosion when the island stood at least 1200 feet higher than at present . . . . The flat valley floor has been produced by slow filling with silt, sand, gravel, secondary volcanic deposits, and probably coral, as the island submerged."

At the foot of ravines and precipices in the upper parts of the valleys, there are slopes covered with talus derived from the surrounding high walls. These slopes vary in width from 100 to 1000 feet, and are inclined 15 to 30 degrees from the horizontal; some approach 40 degrees. The writer believes that these are bedrock slopes veneered with talus, and not cones or fans entirely composed of talus material. Wentworth and Winchell (17, p. 55) make the following statement about similar slopes at the foot of the great cliff (or pali, meaning precipice) on the windward side of the Range: "The base of the pali is flanked by fans of debris, but in many places these apparent fans are only a veneer over buttresses of bedrock which have been left after retreat of the cliff." The talus veneered bedrock slopes in the leeward valleys may be increasing in width at the expense of the cliffs. If this is true, the erosional processes which produced the steep cliffs are no longer as operative as in the past. Blocks and boulders fill the upper courses of the streams, and appear to impede stream flow and hinder downcutting. Coarse gravel chokes the streams, but this may move during excessive rainfall.

Steep walls near the heads of the valleys are serrated by narrow, steep ravines or chutes. The walls of many of these ravines are precipitous; several are true cliffs. Cliffs rim the heads of the valleys, a few being 200 to 300 feet high with 70 to 80 degree slopes. Above these cliffs, hanging valleys open upwards and extend back to the main summit ridges of the Range. The sides of these hanging valleys often have slopes of 45 to 65 degrees. Being in the region of heavy precipitation, the hanging valleys periodically fill with rainwater which cascades as waterfalls down the face of the cliffs below. The courses of these ephemeral waterfalls are outlined perpendicularly on the cliffs by wide zones of black, glistening water-stained volcanic rocks.

J. D. Dana, W. T. Brigham, C. E. Dutton, and C. H. Hitchcock, all early visitors to the island, described many of the topographic features and called attention to the physiographic problems involved in the origin and dissection of Koolau Range. Origin of the leeward theater-like valleys has been a source of controversy for many years. Davis (4, p. 173) observed these valleys in 1914, and stated that the great cliff on the windward side owed its origin to the retreat of weak underlying beds, and that a cliff at the head of "a southward discharging valley" probably owed its origin to the retreat of "the same series of weaker and less resistant underlying beds." Davis accompanied this discussion with a sketch of Manoa Valley (4, p. 174), reproduced here as figure 3.

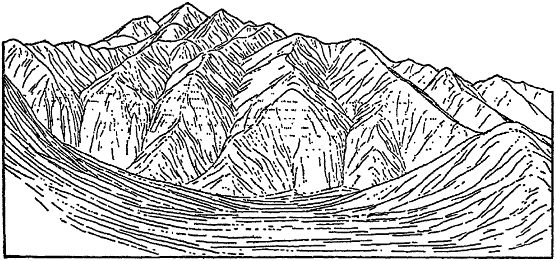

Figure 3. Generalized view of upper valley floor and theater-like head of Manoa Valley, Oahu, Hawaii. Taken from Davis. Highest peak in background is Puu Konahuanui.

Recognizing that conditions differed greatly from those in most temperate humid climates, Wentworth (14, pp. 385-410) believed that "steep-walled, blunt-headed box canyons" were the product of normal processes of erosion for the area and enumerated as principal factors in their formation the porosity of the rocks and their tendency to weather chemically, as well as the climatic factors of high annual temperatures and rainfall. He also suggested that rapid chemical weathering near water table level allowed streams to continue undercutting headwalls and thus determine the configuration of the valleys.

Stearns (12, pp. 24-25) bases his theory for the origin of these valleys upon four conditions: (1) Streams flowing on slopes steeper than three degrees have developed amphitheater-headed valleys; those flowing on gentler slopes have not. (2) Streams undercut nonresistant beds beneath more resistant beds and form waterfalls. The alternating resistant and nonresistant beds usually dip downstream; the waterfalls "increase in height as they follow the dip upstream and tend to coalesce into one high fall." It is intimated (12, p. 25) that the waterfalls are actively incising their canyons in bedrock today. This may be true along the windward slopes of Koolau Range, and for the waterfalls in leeward valleys in the past, but the presence of hydrophytic mosses growing on the face of the cliffs along the course of the falls in the leeward valleys studied suggests the inability of the falls to incise on the face of these cliffs today. (3) Major streams as well as tributary streams erode headward into regions of high altitude and greater rainfall. The tributary streams, increased in number and in vigor, capture headwaters of streams on adjacent sides and enlarge the catchment basins of the major streams near their source, and theater-like valley heads result. The confined water in the dike complex in the interior of the Koolau dome is tapped by the more powerful streams, and this "accelerates stream capture . . . " (4) Captured hanging tributaries plunging over the cliffs produced by the waterfalls remove the narrow ridges between tributary streams by plunge pool action. These narrow ridges also are reduced by landsliding.

Stearns further suggests (12, pp. 25-26) that the valleys are oversized for the volume of water now flowing in them, but it must be rememered that under such conditions no "normal" size has ever been determined for either stream or valley. A considerable amount of talus veneers the bedrock slopes at the foot of the cliffs and occupies the floors of the upper valleys. This suggests that erosion is no longer as active as it was in the past, and that changes have occurred which prevent streams from keeping the upper valley floors clear. If, as Stearns has specified (12, p. 26), the drainage areas above the heads of the leeward valleys have been greatly reduced by piracy from the windward side, and, if general reduction of the height of the Range of approximately 1000 feet by erosion and more than that by submergence has reduced rainfall by one half, then streams flowing in the leeward valleys are much smaller than those that once must have existed. If it can be proven that the talus veneered bedrock slopes are increasing in width, as previously suggested, then the statement that rainfall has been reduced by one half will be substantiated.

Hinds (7) and Jones (8, pp. 55-56) have called attention to these valleys, but did not advance any new explanation for their origin. Cotton (3) has presented a composite review concerning problems involved in the origin of all U-shaped Oahu valleys. He stresses need for objective study of valley sculpture in all tropical rain forest regions.

So little is known of the relative importance of processes of erosion and climatic factors that enter into the formation of these theater-like valleys that the critical reader today is still uncertain of their origin. The correct solution may be determined by conducting further field studies in the Hawaiian Islands and in tropical regions, by study of the literature concerned with similar problems of erosion in tropical climates, and by careful comparison between processes of erosion in the Hawaiian Islands and in other tropical regions.

MOVEMENT OF SMALL PARTICLES ON STEEP SLOPES

The following discussions are based upon many observations in leeward valleys on the movement of small particles of weathered rock debris down through the steep ravines and over the talus veneered bedrock slopes to the main streams on the valley floors below.

The bedrock of the leeward slopes of Koolau Range is highly permeable. It is permeable because of the great number of thin, vesicular lava flows and beds of cinders that form the Koolau dome, and also because of joints, cooling cracks, irregular contacts between flows, and lava tubes and tunnels. The interstices in these volcanic rocks are continually being filled and flushed out with rainwater. This water, after it falls on the surface of the ground, trickles down towards the water table in the rocks beneath. As it passes through the soil, it carries some of the smaller particles of weathered rock with it. Many particles are washed into the cracks and interstices of the rocks below. Some emerge, however, after moving relatively short distances. Those particles that do emerge are deposited on the walls of the steep cliffs and ravines in the leeward valleys. Many are moved farther downhill during subsequent periods of rainfall. Those that emerge out of the rocks on the walls of ravines during a shower either are carried away immediately by the water flowing in the ravine, or, if deposited on the walls above the flowing water, accumulate as thin, reddish brown layers of mud. This mud is deposited as stalactites or incrustations of mud, and veneers the ravine walls. Depending upon shower intensity and volume of rainfall, the interstices of the rocks continue to drain for several minutes or for hours, leaving the ravine walls sheathed with mud. After the water that was flowing through the ravine has drained away, the ravine floor also becomes incrusted with mud. This accumulation of muddy particles makes ascent or traverse of ravines and steep walls not only difficult but dangerous. If the intervals between showers are long, the mud may dry and harden; then travel is easier. Another rainfall, however, cleans away the mud previously deposited; more is dropped again after the next shower. The small particles of weathered rock debris, derived from the high mountain slopes above, are thus removed and transported downward by rainwater—an effective process of erosion.

Movement of rainwater from the cliffs, ravines, and hanging valleys about the heads of the valleys results in the transfer of small particles in the form of a muddy sludge down to the talus veneered slopes below. Quantities of soil and larger rocks from the higher summits above, of course, are also washed down through the ravines to the gentler slopes below. Plant growth is relatively dense on these talus slopes; in some places large blocks and boulders rest on the slopes, but the trees, ferns, and small plants grow up between them. Beneath the plants and decaying masses of humus and forest litter and below the zone of root mats and network of rhizomes, the muddy sludge of tiny rock particles is in continual motion during and for a short time after heavy rainfall. The mud moves downhill through the network of roots, around tree trunks and between the huge boulders that protrude above the decaying forest litter and rotting slimy humus. Travel across such slopes during or immediately after a shower is a slippery process. Occasionally the muddy particles, in the form of mud flows, break out through the forest litter and move down across the forest floor. In this manner, mud flows are spread out on top of the soil and decaying vegetation; new plant growth eventually develops on the surfaces of the mud flows and hinders any further movement. During long intervals between showers, the mud beneath the vegetation becomes dry and hard. Heavy rains soften the hardened mud and movement of the mud continues. If sufficient soil, muddy particles, and decomposed rock debris are carried out from beneath large enough areas of thick matted masses of plants, the forest is thereby lowered onto less decomposed bedrock. Further chemical decomposition will occur, and the surface of solid bedrock beneath the talus thus becomes weathered down to a lower level. Continued removal of overlying soil and weathered rock debris by flow and extrusion of the muddy particles is a valid method of slope erosion. The rate of erosion on these gentler slopes in this fashion cannot compare with that on the steeper slopes above, and proceeds at a slower rate. This flow and creep of small particles of weathered rock debris, however, is a process of erosion that trims off and lowers the talus veneered slopes. Large protruding boulders and rocks also disintegrate by chemical weathering, and are removed particle by particle. Some decayed material is removed in solution.

Perennial streams, flowing on the floors of the upper parts of the valleys between talus veneered bedrock slopes, are hindered somewhat by larger blocks and boulders that have dropped and rolled down from the steep slopes above. Most of the water in these streams is derived from the waterfalls, and from the springs that emerge along the foot of the headwalls and cliffs. A considerable amount of water, however, enters through the sides of the stream channels. This water carries with it the muddy particles that have moved across the adjacent slopes. Most of the particles are transported during heavy showers. Their appearance on the channel sides is not conspicuous during high water periods, as they are quickly swept away. The channel walls continue to drain for several hours after a shower. They become coated with mud as the stream water diminishes in volume and no longer washes off the channel walls. During the next rainy period, the remaining mud is swept away. Accordingly, comminuted debris that has moved across the slopes above is shifted downstream.

Movement of small particles of weathered rock debris, from the higher summits and mountain slopes, down through the ravines and across the cliffs, over the talus veneered rock slopes below to the main streams on the valley floors, is accomplished by these slow processes of soil flow and soil creep. These are important, continual processes of erosion, which operate in tropical and semi-tropical regions, and cannot be disregarded in an inquiry concerning denudation.

AVALANCHING ON STEEP SLOPES

One method of erosion, not overlooked, but which has been neglected by some and considered unimportant by others, is that of landsliding or avalanching. Stearns (12, p. 25) emphasizes landsliding as an active means of erosion which should not be omitted. Wentworth (15) devoted considerable study to the process of soil avalanching with excellent results. He has pointed out its importance as a controlling factor in determination of land forms in semi-tropical mountainous regions. The writer is in full agreement with his observations and conclusions.

Wentworth ably describes abundant evidence of soil avalanching on 40-50 degree slopes; his compilation (15, p. 57) shows "80 per cent of the total between 42 and 48 degrees, with an occasional slope as high as 55 degrees." In addition to this and to the sliding on very steep knife-edge ridges mentioned by Wentworth (15, pp. 61-62), evidence of avalanching was observed by the writer on the 70-80 degree steep slopes at the head of Manoa Valley.

At a distance the steep slopes at the head of the valley present a mottled or spotted appearance. Close inspection reveals that although diversified, the steep slopes have plants in some form growing on them. Some parts of the slopes are mantled with a thick soil and well developed plant society, other parts with thick soil and stunted, scrubby grass growth. Certain other sections maintain no soil at all, but instead are sheathed with yellow-brown mud so thin that layered volcanic cinders can be seen beneath. The soil and vegetation on the slopes are in various stages of development from youthful thin soil to maturely developed thick soil, and old thick soil. No ecologic zoning can be distinguished; distribution of variegated patches of vegetation has no obvious relation to topography or climate.

The writer believes that avalanching occurs on these steep slopes, that maintenance of the slopes has depended upon avalanching as a method of erosion in the past, and that avalanching and mass movement of particles of weathered rock debris control the recession of the steep slopes today.

Wentworth (16) believes that the surface of the bedrock and the base of the detached debris layer, with the presence of rainwater plus the indigenous weathered residuum, should be a surface of movement. He states that this is apparent because of the discontinuity of mechanical properties between the two layers. There can be no doubt that an element of discontinuity exists between the two layers, and that it is of major importance in the movement of debris. In fact, this element of discontinuity should increase when considering the steeper slopes. Avalanching also depends, however, upon the ability or inability of vegetation to aid the soil in adhering to slopes. This factor also increases when considering steeper slopes. The added weight of many plants can easily upset the equilibrium necessary for continual adherence to a steep slope. Certain types of deeply rooted trees can be most tenacious, whereas a mat of shallow rooted grasses and plants responds more readily to the pull of gravity. In addition, the weight of rainwater, suddenly multiplied, can supply the necessary impetus that sets an avalanche in motion. A thick, heavy old soil, which supports only a decadent growth of grasses, is the least tenacious. Mature soil supporting a thick mass of many plants and trees, weakened due to exhaustion of underlying soil, is very susceptible to avalanching. A youthful soil supporting young, luxuriant plant growth is the most adhesive and usually able to withstand every cloudburst.

The variegated patches of plant growth on the cliffs and steep slopes at the head of Manoa Valley and the presence here or absence there of thick soil strongly suggests that intermittent, catastrophic avalanching has occurred. The parts of the steep slopes covered by a thick, old soil which support only grass are locations for immediate avalanching. The areas of mature soil supporting thick masses of plants and trees will avalanche in the distant future. The parts of slopes supporting a thin soil with young, luxuriant plants are locations of past avalanching and of recent soil development. The sections of bedrock sheathed with yellow-brown mud are the recently exposed volcanic rocks. Such alternate, intermittent removal of soil and vegetation and weathered rock is a method of slope recession that appears to be active today.

Recession of steep slopes in the past would certainly have been greatly facilitated by the larger drainage areas above the heads of the valleys, which are now considerably reduced, and by the higher rainfall possible during the periods of emergence and greater height of Koolau Range, as previously discussed. Any further argumentation would be pure speculation at this writing; proof of such erosion for the past will be revealed only by further field work.

CYCLES OF RECURRENT AVALANCHING IN TROPICAL REGIONS

Avalanches occur repeatedly on the face of the steep slopes and cliffs, although recurrent sliding at any one place is dependent upon many factors. These factors should include the rapidity of soil development and the establishment and growth of plant societies, the speed of soil exhaustion and decay of plant remains, and the instability of the weakened mass of soil and plant debris. A cycle of recurrent avalanching is tentatively suggested, briefly, as follows (1) fresh bedrock exposed; (2) the establishment and succession of mosses, plants, ferns, and trees to the culmination of maximum plant relationships possible, with accompanying soil development to great depths; (3) long and continued use of all available soluble minerals and organic matter; (4) final exhaustion of the soil mantle with deterioration of the plant societies, and alteration to a degenerate growth of grasses; (5) desiccation and hardening of the mass of soil, decaying litter, and rotting humus; (6) instability of the mass of debris produced during prolonged rainfall; and (7) catastrophic avalanching exposing fresh bedrock slopes again.

If the thickness of soil at the time of avalanching and the time required for one cycle of recurrent avalanching could be determined, a rough estimate for the rate of recession of such steep slopes could be given. Wentworth (15, pp. 62-63) estimated that in a period of eight years in an area of about 15 square miles, in that part of the Honolulu watershed best known to him, an equivalent of 200 slides involving an acre each had occurred. He also estimates (15, p. 63) that an average thickness of one foot of material is removed by each slide, and that if the material is carried out of the area, that would be "equivalent to a rate of about 1 foot in 400 years for the whole area." Freise (6) in his study of Brazilian tropical rain forests postulated similar cycles of recurrent avalanching. The avalanching he describes occurs on steep slopes and cliffs of granite and gneiss. He has realized that he, too, is limited by the lack of precise knowledge as the writer, for he states (6, p. 155): "When numerical values for the progress of weathering, for erosion by disintegration, and for the rise and decline of ever-moist tropical primeval forests . . . are determined, then a figure for rate of erosion can . . . be postulated." (S.E.W., free translation) Freise's estimates for the rate of recurrent avalanching, however, of about 300-400 thousand years, do not seem to be based upon any statistical data, and cannot be considered useful at this time.

In 1940, Bryan (1) called attention to the work of Thorbecke (1927), Jaeger (1927), and Sapper (1935),[2] as well as to that of Freise (5; 6), on the retreat of slopes in tropical regions. The writer has been unable to study the works of Thorbecke or Jaeger. Sapper describes in detail processes of soil flow and soil creep similar to those that occur on the steep slopes of Oahu. Freise (5) presents his observations in the coastal mountains of Brazil with a complete description of the tropical forest cover and its effect upon the supporting soil. The processes of soil flow and soil creep that he sets forth are similar to those described by Sapper and to those presented above for Oahu. Freise also recognizes (5, p. 91) that soil flow and creep due to the delayed effects of rainwater in reaching the soil by way of tree trunks and their roots is a different process of soil flow than that recorded above. In 1938, he first describes the form, size, construction, and environment of "inselberge," then analyzes their origin meticulously as they developed both in the dry regions of northeast Brazil and in the constantly moist tropical coastal forests. It was in this paper (6, p. 154) that he advanced the doctrine of "erosion induces deforestation" (Erdfliessen Entwaldung schafft), and introduced a cycle of recurrent avalanching for such moist tropical regions. His cycle (6, p. 155) is summarized briefly, as follows:

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[2] Taken from Bryan, p. 268:

Thorbecke, F.: Der Formenschatz im periodish trockenen Tropenklima mit überwiegender Regenzeit. *Morphologie der Klimazonen* (ed. by Thorbecke). *Düsseldorfer geogr. Vort. u Erört.* Breslau, 1927.

Jaeger, F.: Die Oberflächenformen im periodish trockenen Tropenklima mit überwiegender Trockenzeit. *Morphologie der Klimazonen* (ed. by Thorbecke). *Düsseldorfer geogr. Vort. u. Erört.* Breslau, 1927.

Sapper, K.: Geomorphologie der feuchten Tropen. *Geogr. Schriften,* 7, 154 p., 1935.

> "exposed granite (or gneiss) ridges — humus formation (upon a covering of lichen and small plant growth) — immigration of trees (probably leguminous at first) — development of a great primeval forest on a large scale with a combination of carbonic acid and humous acid type of decomposition — decay phenomena with soil flow and soil incrustations — atrophy of the forests to a deterioriation forest — domination and conquest by the grassy growths, destruction of the impoverished soil residue — emergence of the vegetation-freed ridges again (S.E.W., free translation)."[3]

When examined in detail, this cycle of recurrent avalanching is quite similar to that suggested for the steep slopes on Oahu. Differences in the rate of decomposition of the two types of underlying rocks, volcanic rocks in Hawaii and granite and gneiss in Brazil, make it difficult to compare what appear to be similar processes of erosion in nearly similar climatic regions. Freise's cycle, however, is much more complicated and involves a greater number of special conditions than does the one postulated for the Hawaiian Islands. Correlation between the two areas should not be attempted at present.

CONCLUSIONS

The movement of small particles of weathered volcanic rocks from high mountain summits to the valley floors by soil flow and soil creep has been recorded as observed in certain leeward valleys of Koolau Range, Oahu, Hawaii. Emphasis has been placed upon this form of erosion as a valid and important corollary to the more obvious methods of stream abrasion, sliding of debris and avalanching of soil. The processes of soil flow and creep are not manifest, and need to be examined and studied during their spasmodic occurrence. The avalanching on 70-80 degree slopes in one leeward valley is advanced as supplemental to Wentworth's excellent presentation of analogous features on 40-50 degree slopes. A cycle of recurrent sliding, similar to that claimed by Freise for steep slopes in the wet tropical coastal regions of Brazil, is suggested for Oahu, and includes the development of soil on 70-80 degree bedrock slopes with establishment and growth of plants, later soil exhaustion after maximum consumption, instability of soil and decaying plant debris induced by rainfall, and avalanching to expose bedrock again. Figures for rates of recurrence, especially those of Freise, cannot be considered at this writing.

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[3] Freise did not mention avalanching in this summary, probably because he discussed it at such great length in his text, but his intent is obvious.

With the exception of that of Freise, precise, exacting work on the erosion of slopes in tropical regions is rare. No one has come forward in the literature to attract attention to the mechanics of movement of debris on slopes in tropical regions, as, for example, has Walther Penck for humid regions. The need, however, for such an endeavor is evident; additional, accurate, discriminating field work is essential.

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THE POLAROGRAPHIC METHOD FOR DETERMINING TRACE ELEMENTS IN ROCKS AND MINERALS

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ABSTRACT. Polarographic analysis is based on the measurement of current developed in a solution contained in an electrolytic cell with a dropping mercury electrode. The magnitude of the current is proportional to the concentration of the element, and its measurement establishes quantitative determination. Identification of the element is made from the "half-wave potential," each electro-reducible or electro-oxidizable substance having a characteristic value.

The method is particularly suitable for quantitative analysis of amounts in the range of 10^{-2}M to 10^{-6}M (about .X% to .000X%), for many metals and nonmetals occurring in rocks and minerals. It offers greater sensitivity for determining Cu, Zn, Cd, and possibly other elements, than spectrographic analysis (unless special techniques are used), although it lacks the latter's advantage of multiple determinations from a single spectrogram of very small samples. Usually, not more than three or four elements are determined simultaneously with the polarograph.

In many cases, determinations can be made in the presence of other elements, which by ordinary chemical methods are impossible or impractical. Determination of many major constituents can also be made, although with less accuracy, thus permitting a quick, simple and approximate analytical method suitable for broad surveys. Outstanding applications are: simultaneous determination of ferrous and ferric Fe; Ba in the presence of Sr, Ca and Mg; Mo in ores of Mo; Cb in the presence of Ta and Sn. About forty elements are determinable.

INTRODUCTION

ALTHOUGH not well known in geochemical circles, polarography was originated about twenty-five years ago by Professor Heyrovsky, at Charles University in Prague. The analytical method is based upon the interpretation of current-voltage curves of solutions in an electrolytic cell. Under certain conditions, these curves lead to the identification and quantitative determination of one or more substances present in the solution.

The polarograph is a precision instrument which automatically records the current-voltage changes taking place in an electrolytic cell containing a dropping mercury electrode. There are three types of polarograph: Heyrovsky's original design, providing a photographic record of galvanometer deflections; a second type in which a mechanical pen produces a continuous chart record of current-voltage changes; and a third type in which a cathode-ray oscillograph acts as the recording device. The model with which data for this paper were obtained is an improved pen-recorder type (pl. 1) designed by J. H. Schulman, H. B. Battey, and D. G. Jelatis (1947) at Massachusetts Institute of Technology.

A highly simplified circuit diagram of the polarograph is seen in figure 1. When a regulated voltage supply from a storage battery is applied to the terminals of the cell containing a dropping Hg electrode, an electric current flows through the test solution in the cell, which is amplified and recorded on the milliammeter chart.

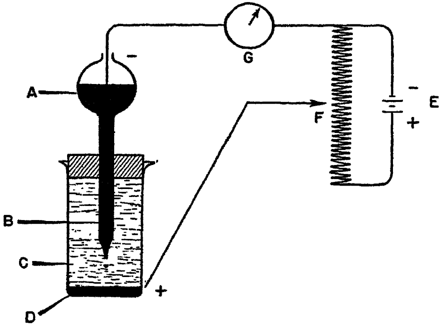

Figure 1. Schematic circuit diagram of the polarograph. *A*, Hg reservoir; *B*, dropping Hg electrode; *C*, test solution; *D*, large quiet Hg electrode; *E*, storage battery; *F*, voltage regulator; *G*, recording device.

Any electro-reducible or electro-oxidizable substance produces a characteristic curve under these conditions. Most of the metallic and nonmetallic elements occurring in rocks and minerals, as well as many organic substances, can be determined. The method is especially suitable for small concentrations, and as very little solution is needed (as little as .005 – 1.0 ml. with microcells, although 10 – 20 ml. ordinarily), mere traces can be measured quantitatively. At the Naval Research Laboratory, we have used the polarograph for determining traces of Pb, Mn, Tl and Sn in calcite and halite; and have also analyzed synthetic sphalerites for Zn and Cd as major constituents, although with less accuracy.

BASIC PRINCIPLES

The heart of the polarograph is its electrolytic, polarizing cell (fig. 2) containing a dropping Hg electrode. This electrode is a length of glass tubing with a fine capillary opening, through which mercury from a reservoir drops slowly into the solution. The dropping electrode is generally employed as the cathode (for electro-reduction) but it is also possible to reverse the polarity of the cell and employ it as anode in the study of electro-oxidation reactions. The other electrode is a large quiet pool of mercury at the bottom of the cell, or an external, saturated calomel half-cell. The two electrodes are connected to a storage battery in such a fashion that a slowly increasing voltage can be gradually applied to the cell terminals. When voltage is applied to the cell, a current develops due to electro-reduction or electro-oxidation of the substance in solution.

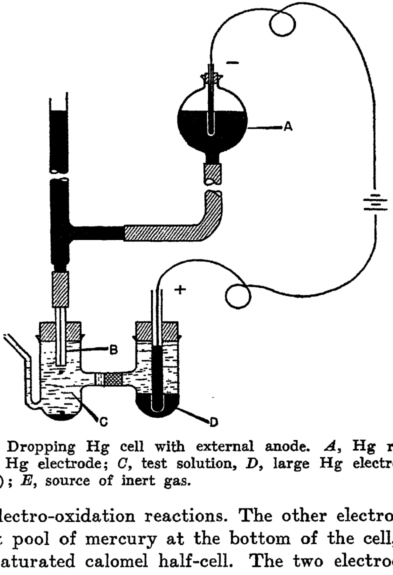

Figure 2. Dropping Hg cell with external anode. *A*, Hg reservoir; *B*, dropping Hg electrode; *C*, test solution, *D*, large Hg electrode (external anode); *E*, source of inert gas.

At first, with only a small voltage supplied, the current is exceedingly small. This is the "residual current" (see fig. 3), due to minute traces of impurities always present and to capacitance effects. As the voltage is increased, a point is reached (the "decomposition potential") where there is now sufficient energy for the electro-reduction (or oxidation) reaction in the cell to take place, and a sizeable current develops rapidly. The magnitude of the current is actually very small, rarely exceeding 50 microamperes. Finally, the current levels off, reaching a limiting value, and further increase in voltage has no effect. The net increase in current (the difference between the "limiting current" and the "residual current"), other factors being constant, is directly proportional to the concentration of electro-reducible (or oxidizable) substance in solution, and is the basis of quantitative measurement.

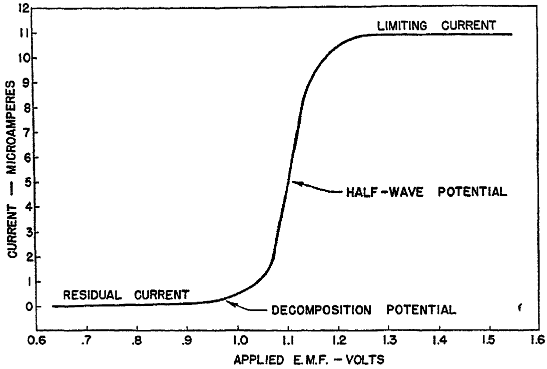

Figure 3. Typical current-voltage curve.

In order to obtain a limiting current, which is caused by an extreme state of concentration polarization (at the dropping Hg electrode), that electrode[1] must be very small and the other, relatively large. During the course of the electrolytic process, the concentration of the reducible (or oxidizable) substance is decreased near the surface of the mercury drops and, by diffusion, a fresh supply of reducible substance from the body of the solution continually reaches them. As the voltage is increased, the concentration of reducible substance at the surface of the mercury drops soon becomes so small as compared with its concentration in the bulk of the solution, that the difference approaches a constant value, and the rate

[1] A rotating, solid platinum micro-electrode is sometimes substituted.

of diffusion also becomes constant. In this sense, the limiting current is described as a "diffusion current."

The decomposition potential is characteristic of the electroreducible (or oxidizable) substance present in solution, but is also a function of its concentration, hence it cannot be made the basis of qualitative identification. Instead, the "half-wave potential" (see fig. 3) of a given substance is used for identification. This is the voltage point on the curve where the net current increase has reached one-half its limiting value, and it is independent of the concentration.

LABORATORY TECHNIQUE

The presence of a large excess of an indifferent compound or "supporting electrolyte" (one whose ions do not participate in the reaction) is necessary to ensure a limiting current that is truly proportional to the concentration. The best supporting electrolyte for a particular analysis depends largely on the substance to be determined. It is often convenient to treat a major constituent as the supporting electrolyte, such as NaCl in the analysis of halite or $CaCl_2$ for calcite (obtained by dissolving calcite in HCl). Factors to be considered in selecting a supporting electrolyte are: (1) voltage range to be covered; (2) possible formation of complex ions; (3) secondary reactions with mercury; (4) possible presence of interfering elements; (5) concentration of supporting electrolyte required (usually ranges from 0.1M to 1.0M).

Table I lists some common supporting electrolytes.

TABLE I

Common Supporting Electrolytes

1. HCl, H_2SO_4, HNO_3, $HClO_4$
2. NaCl, KCl, $CaCl_2$, etc.
3. NaOH, KOH, $NH_4OH + NH_4Cl$
4. Tartrates, citrates, oxalates
5. KCN, KCNS
6. Tetramethyl ammonium hydroxide
7. Pyridine + pyridinium chloride

An actual chart record, or polarogram, made with the Schulman-Battey-Jelatis instrument is shown in figure 4. The test solution contained equal amounts of Pb^{++} and Cd^{++}. When several substances are present in the solution, each one will produce its own characteristic wave on the curve, provided that the half-wave potentials of these substances are not too close together. The over-all spread, or the height, of each

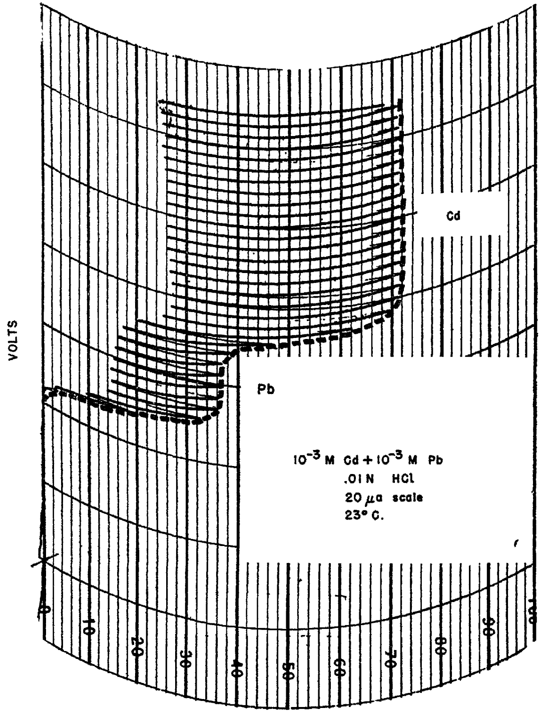

Figure 4. Polarogram of Pb and Cd present in the same solution; simultaneous recording. The heavy broken line edging the peaks has been drawn in to delineate the shape of the current-voltage curve.

wave (indicated by the horizontal portions of the outlined curve in figure 4) is measured to determine the amounts of each

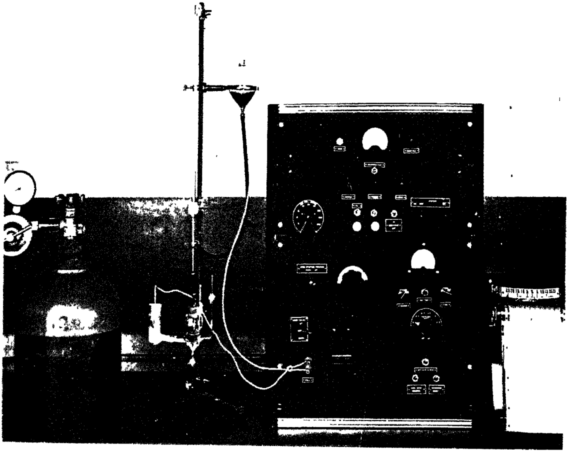

Plate 1. Naval Research Laboratory model, an improved pen-recorder type of polarograph. Reading from left to right: tank containing inert gas (nitrogen), dropping Hg cell with external anode, polarograph, recording milliammeter.

substance present. The horizontal oscillations of the polarogram (see fig. 4) merely indicate the growth and breaking

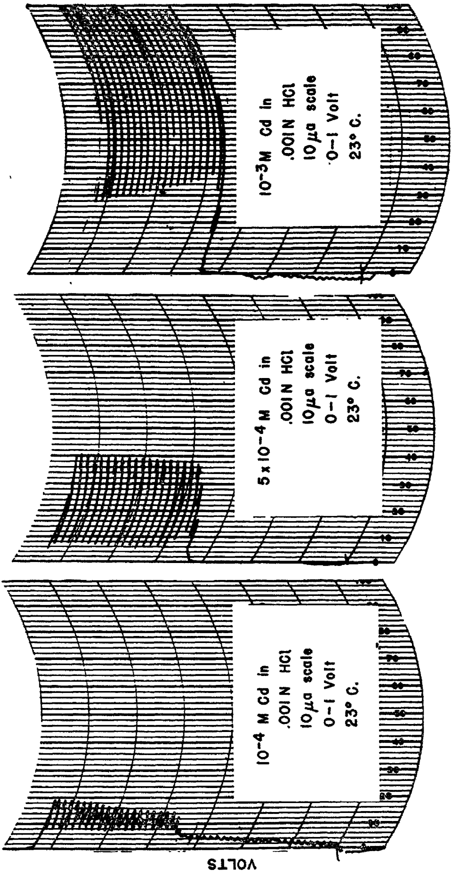

Figure 5. Polarograms of Cd, showing waves produced by different concentrations, increasing from left to right.

away of each mercury drop, and are ignored in making measurements.

In figure 5 are polarograms of Cd solutions of three different concentrations — 1×10^{-4}M, 5×10^{-4}M, 1×10^{-3}M (.01, .05, .1 g. per liter, respectively). The effect of concentration on the magnitude of the wave, or limiting current, is readily seen.

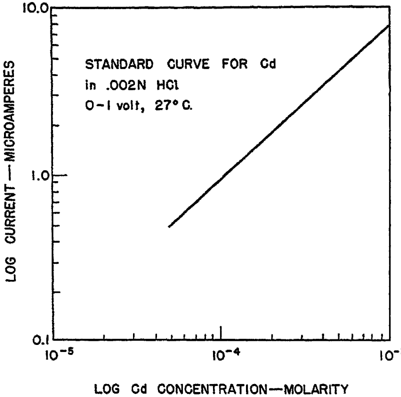

Figure 6. Standard Cd curve, for computing Cd concentration on the basis of recorded polarograph current.

More or less pronounced humps, or "maxima," are seen on polarograms of some substances. These tend to distort or increase the effect of the limiting current and make quantitative measurement impossible. While the origins of such maxima are complex, in practice it is usually simple to suppress them by the addition of a trace of gelatin, glue or certain dyes.

An analysis is rather simply and quickly made. The sample to be studied is suitably dissolved; the solution is placed in the cell; free oxygen (dissolved air) is always removed, by bubbling through an inert gas such as nitrogen; the current-voltage curve is recorded; the curve is then compared with standard curves (fig. 6) of known concentrations in a comparable type of solution; and the concentration of the unknown in the test sample is calculated.

Once standard curves have been prepared, these can always be referred to with reliability, provided the same dropping electrode has been employed. Temperature corrections may be necessary unless the cell is set in a constant temperature bath, or operated in an air-conditioned room. There is a 2% increase in the magnitude of the current per centigrade degree rise in temperature.

The composition of a solution about to be analyzed must be such that the ion to be measured is maintained in a definite valence state. Moreover, the wave produced when the reaction occurs must not be obscured or distorted by other nearly simultaneous reactions. For this reason, it is sometimes necessary to remove or suppress by chemical means any interfering elements. The removal of oxygen, because of its readily-developed, large wave masking other waves in that voltage range, is a case in point. It is usually sufficient to remove only the bulk of the interfering substance. Furthermore, precipitates, if formed in the process of separating out interfering elements, need not always be filtered off but can remain at the bottom of the solution, unless present in large amounts.

On the other hand, it is often possible to avoid such interferences by use of a specially selected, supporting electrolyte in the solution, or by various adjustments of the instrument itself. For example, the half-wave potential of Pb is only .06 volt more positive than that of Tl, and the two waves nearly coincide. There must be at least 0.1 volt difference between the half-wave potentials of any two substances for satisfactory wave resolution. By adding a large amount of potassium tartrate to the solution, the Pb wave is shifted away by several tenths of a volt, while the Tl wave remains unchanged. This allows for the reliable determination of Pb and Tl in the same solution simultaneously.

The polarograph permits the simultaneous estimation of several metals—sometimes as many as five or six, usually not more than two or three. Thus, where only small samples are available for complete analysis, spectrochemical examination is advisable.

GEOCHEMICAL APPLICATIONS

Geologists and geochemists unfamiliar with polarographic analysis may, at this point, jump to the conclusion that the complex chemical nature of most rocks and minerals would cause such serious interferences as to negate the value of the method. This is not so. In Table II are listed some of the more outstanding possible applications of polarographic analysis to geochemistry. Included are certain determinations which are impossible or impractical by wet chemical methods.

TABLE II

Polarographic Determinations Applicable to Geochemical Problems

1 — Tl in presence of Pb.
2 — Pb and Cd in presence of excess Zn.
3 — Sn in presence of Cb and Ta.
4 — Cb in presence of Ta and Ti.
5 — Sb, Bi, Pb, Cd in presence of excess As.
6 — Bi in presence of excess Pb, Cd, Sb.
7 — Bi and Sb, simultaneously.
8 — Zn and Cd, simultaneously.
9 — Al, Cr and Fe-ous, simultaneously.
10 — Fe-ous and Fe-ic, simultaneously.
11 — In in presence of Zn.
12 — Ni in presence of large excess of Co, Cu, Zn.
13 — Ni and Zn, simultaneously.
14 — Co in presence of excess Ni.
15 — Rh in presence of Ir and Pt.
16 — U in presence of Th, V, Cr.
17 — *Ti in Ti-minerals (rutile, yttrotitanite).
18 — *Mo in ores of Mo.
19 — W in presence of Mo.
20 — Se in presence of Te.
21 — Se and Te, simultaneously.
22 — *Na + K in lake waters, soils, Al-silicates.
23 — Ba in presence of other alkaline earths.
24 — *Iodate traces in nitrate (Chili saltpetre).

* — Procedures already in use.

Zeltzer (1932) worked out a simple procedure for determining Ti in minerals such as rutile or yttrotitanite. The mineral is fused with $KHSO_4$ in a platinum crucible, the melt is dissolved in dilute H_2SO_4, the solution is then transferred to the cell and the polarogram recorded. Al and ferrous-Fe do not interfere, but Cr may.

Kanevskii and Shvartsburd (1940) determined Mo in ores containing from .005% to over 50% Mo, with an accuracy of 2–5%. The ore is fused with alkali flux; leached with hot water; Mn removed by peroxide treatment; H_3PO_4 is then added to an aliquot portion of the solution and the polarogram is recorded.

The polarographic waves for Na and K are so close together that only the sum of the two can be obtained. Sometimes the Na is later determined chemically with zinc uranyl acetate, and K calculated by difference. Majer (1933) developed a simple procedure for determining Na + K[3] in aluminum silicates. A very small sample is decomposed with H_2SO_4 and HF; the dried residue is treated with a drop of H_3PO_4 to precipitate Fe, Al, Ca, etc.; and tetramethyl ammonium hydroxide is added. The material, precipitate and all, is transferred to the cell, and the polarogram is recorded. This method gives an accuracy of about 5%, where the amount of Na + K averages 2% of the total. However, a complete analysis can be made in about thirty minutes, so the method may be particularly useful for broad surveys, where time and simplicity of operation are important factors.

Of the alkaline earths, only Mg cannot be determined directly at present (its wave is poorly defined). Ba and Sr, however, can be easily determined in the presence of relatively large amounts of Ca and Mg. Carruthers (1943) has developed an ingenious procedure for the indirect determination of very small amounts of Mg. The sample is treated with 8-hydroxy quinoline to precipitate Mg in complex form. The precipitate is dissolved in HCl and added to a buffer solution. The quinoline wave is recorded polarographically and then the equivalent amount of Mg present is calculated.

Cb can be determined quite readily, according to Stromberg and Reinus (1946), in the presence of Ta and Ti, by fusing the sample with K_2CO_3 and dissolving the melt in HNO_3. Sn can also be determined in the presence of Cb, Ta and Ti.

ACCURACY AND SENSIVITY

Under ordinary working conditions, the accuracy of the polarographic method is of the order of 2–5% in the usual concentration range of 10^{-2}M to 10^{-6}M, which is roughly equivalent to a range of tenths to ten-thousandths of 1%. Considering the small amounts actually involved, such an error is not unsatisfactory and compares most favorably with that of other analytical methods.

In fact, the polarograph offers greater sensitivity than the spectrograph (Claffy, 1947), unless special techniques are used, with Cu, Zn, Cd and possibly some other elements. As with the spectrographic method, the sensitivity varies somewhat from element to element.

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[3] More exactly, this value may be the sum of all the alkali metals present, since their half-wave potentials are all rather close together.

TABLE III

Periodic Chart of the Elements Determinable by Polarographic Methods‡

H																	**He**
Li	**Be**											**B**	**C**	**N**	**O**	**F**	**Ne**
Na	Mg											Al	**Si**	**P**	**S**	Cl	**A**
K	Ca	Sc	Ti	V	Cr	Mn	Fe	Co	Ni	Cu	Zn	Ga	Ge	As	Se	Br	**Kr**
Rb	Sr	**Y**	Zr	Cb	Mo	**Tc**	**Ru**	Rh	**Pd**	Ag	Cd	In	Sn	Sb	Te	I	**Xe**
Cs	Ba	**La***	**Hf**	**Ta**	W	Re	**Os**	**Ir**	**Pt**	Au	**Hg**	Tl	Pb	Bi	**Po**	**At**	**Rn**
Fr	Ra	**Ac**	**Th**	**Pa**	U	**Np**	**Pu**	**Am**	**Cm**								

* and rare earth elements.

‡ Elements in **Bold** type cannot be determined,

Of the 96 known elements (Table III), about 40 can be determined polarographically at present. To the writer's knowledge, the following elements of interest to geochemists are not determinable by the polarographic method: Si, P, S, Ge, Be, Th, Ta. Insofar as geochemical applications are concerned, the method is still in its infancy. Much systematic investigation must be carried on to develop practical analytical procedures. With further research will undoubtedly come improved procedures to eliminate interference and to extend the method to conditions now considered beyond its scope.

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Washington, D. C.

DISCUSSION AND COMMUNICATIONS

CHAMBER'S MINERALOGICAL DICTIONARY

In Dr. Horace Winchell's review of *Chamber's Mineralogical Dictionary* in the December number of the JOURNAL he states "Acknowledgement of the names of museums or collections containing the specimens illustrating the dictionary would be useful . . ." I should like to call attention to the fact that these illustrations appear to be taken from L. J. Spencer's "The World's Minerals" published by Frederick A. Stokes Company in 1911. In the preface to Dr. Spencer's book he states: "The forty colored-plates have been prepared under the supervision of Dr. Hans Lenk, Professor of Mineralogy and Geology in the University of Erlangen, and many of the pictures represent actual specimens belonging to the mineral collection under his charge." It would seem also from the statement of Spencer that some of the pictures of mineral specimens were not from actual minerals but were, instead, drawings.

EDWARD S. SMITH

INTERNATIONAL CONFERENCE ON QUATERNARY RESEARCH

The Fourth International Conference of the International Association of Quaternary Research (INQUA) is to meet in Budapest late this summer, from August 22 to September 15. Students in all phases of Pleistocene research — stratigraphy, anthropology, paleontology, geomorphology — are cordially invited. There will be two excursions, each approximately a week long, one to Western Hungary (Late Balaton, the Danube above Budapest), one to Eastern Hungary (eastern Hungarian plain, northern bordering mountains).

Direct circularization of interested persons has been greatly hampered by the lack of adequate lists of names, but it is hoped that as many as can will attend. Those planning to attend are requested to inform the

Hungarian Committee for INQUA IV
Hungarian Geological Institute
XIV Vorosilov-út 14
Budapest, Hungary

from whom full particulars can be obtained.

REVIEWS

Geological Explorations in the Island of Celebes; Geological Summary and Petrology; under the leadership of H. A. BROUWER. Pp. 346; 25 Pls., 25 figs., 2 fold-in maps. Amsterdam, 1947 (North Holland Publishing Co., 27.50 guilders, appr. $10.40).—This volume is a second report on the results of a geological expedition into central Celebes conducted in 1929. The first report, published in 1934 (Geologische onderzoekingen op het eiland Celebes; Verhandelingen van het Geologisch-mijnbouwkundig genootschap voor Nederland en kolonien, geol. serie, X, pp. 39-218, 1934; not seen), included a general geological summary and reports on the fossils collected. The present report is written in English.

The bulk of the present volume consists of reports on detailed petrologic studies of the rocks collected by the expedition. These reports, by W. P. de Rouver and C. G. Egeler, deal largely with the metamorphic rocks. The non-metamorphosed igneous rocks are also described though in less detail; apparently the sedimentary rocks were not studied. The rock descriptions are fairly full but not quantitative. The detailed information is well organized and is lucidly generalized on the basis of modern theories of metamorphic facies.

Preceding the petrologic reports is a new summary by H. A. Brouwer of what is known of the geology of Celebes, incorporating the new petrologic results. Unfortunately this summary is not well organized, and no clear picture of present geological knowledge about Celebes is presented. The reader wishing to learn what general results have come from the expedition must dig them out for himself, at the grave risk of making incorrect generalizations because of his lack of familiarity with the area. Professor Brouwer would have greatly enhanced the value of this work if, in the light of his great knowledge of the area, he had drawn what generalizations he could, even though they would naturally require much qualification because of the meagre data. A general map of Celebes, showing the location of the many places mentioned outside the area of the detailed maps, would also have been very helpful.

The area studied lies in central Celebes at the junction of the four great peninsulas that form most of the island (Figure 1). These four arms form two roughly concentric arcs convex northwestward. The two arcs have quite different geology and appear to represent the two geanticlinal belts which can be traced, according to several geologists, from Burma to the Banda Sea and thence through Celebes into Mindanao. The area described in this volume is apparently the only place in Celebes, or for that matter in the

East Indies south of Mindanao, where the junction between the two belts is now above water; hence its great importance and interest to students of tectonics.

The rocks of the eastern arc include a group of metamorphic rocks, now in the epidote-amphibolite facies, and a series of sedimentary rocks which were not affected by this metamorphism and are presumably younger. The sedimentary rocks include much limestone, some carrying *Globigerina,* some varicolored, and some strongly bituminous, and also considerable radiolarian chert. These rocks suggest deep water, but shallow water deposits are also known, including much alternating shale and sandstone (of "flysch" type, according to Brouwer), and also sandy limestone and coralliferous limestone. The deep water deposits range in age at least from Permian to middle Cretaceous; the shallow water deposits from Late Triassic to Early Cenozoic. A reader unfamiliar with Celebes cannot ascertain if the known distribution of these two facies makes any intelligible pattern.

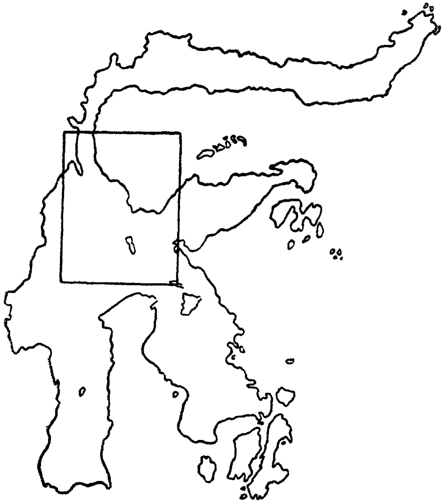

Figure 1. Outline map of Celebes, showing area investigated by Netherlands Indies Geological Survey in 1929.

Volcanic rocks are reported from only one area in eastern Celebes, where they are at least pre-Pliocene. However there are large bodies of ultrabasic and basic intrusive rocks ranging from peridotite (and serpentine) to quartz diorite and including many albitic types. The age of these intrusives is not certainly known, but they are reported to cut Cretaceous rocks and may be Early Cenozoic. Both igneous and sedimentary rocks are severely deformed. Although the general strike is roughly parallel to the trend of the arc, observations to date do not permit of generalizations about the force couple involved.

In central Celebes, rocks like those of the eastern arc continue westward toward the western arc, but they show progressively more metamorphism westward. This metamorphism is quite distinct from the epidote-amphibolite metamorphism of the eastern arc, on which it is in places superposed, and it affects also the sedimentary and igneous rocks that escaped the earlier metamorphism. In the lower grade eastern part of the metamorphic belt the rocks are phyllite and muscovite schist associated with some crystalline limestone and quartzite, and some of the rocks are characterized by alkalic pyroboles (esp. glaucophane) and lawsonite. In the higher grade western part of the belt, the rocks approach biotite gneiss and are characterized by potash feldspar.

Along the western side of this metamorphic belt is a fault trough called by Brouwer the median zone. It contains unmetamorphosed sedimentary and volcanic rocks of Late Cenozoic age which rest unconformably on the metamorphic rocks but are themselves strongly folded. West of the fault trough and separating the rocks hitherto described from the rocks of the western arc of Celebes is a "median line," a narrow zone of mylonite dipping steeply westward. How Brouwer interprets this line, and what its relation is to the Late Cenozoic deposits in the adjacent fault trough, is not clear from the text. The reviewer suggests that it may represent a great thrust fault, separating the rocks of the eastern and western arcs.

The predominant rocks west of the median line are metamorphic rocks intruded by large batholiths, chiefly of granodioritic composition. Though basic and ultrabasic intrusions also occur, they are minor in amount compared both to the granodiorite and to the ultrabasic rocks of the eastern arc. The metamorphic rocks show evidence of two stages of metamorphism: an older regional metamorphism chiefly to the staurolite-amphibolite and epidote-amphibolite facies, and a younger metamorphism associated with the batholiths and characterized by andalusite and sillimanite. In addition all the rocks show a late cataclastic deformation. The batholiths are also held responsible for the potash feldspar gneiss and glaucophane schist metamorphism east of the median line. As the later of the two metamorphisms on each side of the line are thus correlated, Brouwer and Egeler suggest that the older amphibolitic metamorphisms are also to be correlated. Some rocks are reported in the western arc which appear to show only the effects of the later metamorphism.

Also present in the western arc are unmetamorphosed sedimentary rocks, chiefly sandstone, conglomerate, and shale, associated with thick volcanic and pyroclastic rocks. These rocks appear to range in age from Late Cretaceous or Eocene to Late Cenozoic. Their relation to the metamorphic rocks is not clear, but some low-grade metamorphic rocks of Eocene age are also present. Volcanics as late as Pleistocene are reported in west-central Celebes, and there are still active volcanoes at the northeast tip of the western arc. By contrast recent volcanics are unknown in the eastern arc.

If the proposed correlation of the metamorphic epochs is correct, a general geologic history can be deduced. The oldest rocks are those exhibiting the amphibolitic metamorphism; in the eastern arc this metamorphism appears to be pre-Mesozoic, but in the western arc its age cannot be determined. The next series of rocks ranges from unmetamorphosed sedimentary rocks of Mesozoic and perhaps Early Cenozoic age to the east to highly metamorphic gneisses to the west. In the east these rocks are intruded by large bodies of ultrabasic and basic igneous rocks, in the west by granodiorite batholiths. The orogenic episode represented by the metamorphism and at least the acid intrusives was mid-Cenozoic (alpine according to Brouwer). Judging from a diagram presented by Brouwer, the median line (and probably the late cataclastic deformation) were also produced at this time, though he nowhere discusses the matter. Since this orogeny, Upper Cenozoic sediments have been deposited in a number of separate basins, notably a fault trough close to the median line, and there has been great volcanic activity along the western arc. Late Cenozoic deformation occurred in the median zone and also in the southwestern peninsula of the island and perhaps elsewhere.

Brouwer compares the structure thus revealed with that of southwest Japan, where there is also a median line separating igneous and metamorphic rocks on the northwest from lower grade metamorphic rocks on the southeast (see this JOURNAL, vol. 246, pp. 57-61, 1948), and he points out both the resemblances and the differences, the latter being especially in age relations. However, one interesting implication of this comparison remains undiscussed. The older interpretation of the median tectonic line of Japan saw it as the join between two orogenic belts of different ages, but more recently it has been considered as a great thrust fault in the core of a single orogenic belt. If the comparison with Celebes is sound, then the two geanticlinal belts that meander through the East Indies do not have quite different orogenic histories, as has generally been assumed, but are parts of a single orogenic belt, affected as a unit by the last major deformation. If so, it is amusing to note that the two belts cannot even be neatly ticketed as eugeosyncline and miogeosyncline, for one has deeper water deposits and ultrabasic intrusions, the other batholithic granitic intrusions and late volcanic activity.

The above summary of the tectonics of Celebes, extracted from Brouwer's report, is undoubtedly seriously defective in many respects, owing to the reviewer's unfamiliarity with the region and his lack of knowledge about the interrelations of the many isolated facts reported. The ideal man to prepare an adequate summary is of course Professor Brouwer, and it is greatly to be regretted that in the present volume he has not done so. JOHN RODGERS

Illustrated Catalogue of North American Devonian Fossils: Unit 1-E, part a, Family Auloporidae; edited by ERWIN C. STUMM.—This, the first unit of the catalogue to be devoted to corals, includes 114 cards and covers 91 species of tabulates assigned to 9 genera. The most prolific genera are *Aulopora, Drymopora* (which name takes precedence over *Ceratopora*), and *Syringopora.* The inclusion in the family Auloporidae of the genera *Romingeria* and *Syringopora* seems to indicate that the editor would suppress the family Syringoporidae and include the genera heretofore assigned to it in the family Auloporidae.

This unit of the catalogue is more than a mere assemblage of original descriptions and figures. The editor has revised the description where needed, in many instances has added supplementary figures, and has given the species their proper generic assignment according to present understanding.

This group was originally assigned to Drs. C. L. and M. A. Fenton, who prepared about two thirds of the cards, but the task has been completed by Dr. Erwin C. Stumm, who is now in charge of the coral group as a whole. Unit 1-E, part a, is a most welcome contribution to the Devonian fossil catalogue.

CARL O. DUNBAR

A Monograph of the Existing Crinoids; Vol. 1, The Comatulids, Part 4b, Superfamily Mariametrida (concluded—the family Colobometridae) and Superfamily Tropiometrida (except the families Thalassometridae and Charitometridae); by AUSTIN H. CLARK. U. S. National Museum Bulletin 82, 1947—This is the fifth large quarto volume to appear as a part of "volume 1" of Dr. Clark's great monograph on the existing crinoids. Its arrangement is strictly taxonomic, continuing where part 4a left off, but its scope is far more than a mere description of genera and species.

Following a specific description these are commonly notes on peculiarities of color variation, abnormalities of form or distribution, parasites, history of discovery or other pertinent observations. The geographic and bathymetric range is indicated in all cases where it is known. The volume is a treasure house for those especially interested in existing comatulid crinoids.

CARL O. DUNBAR

The Corridor of Life; by W. E. SWINTON, illustrated by ERNA PINNER. Pp. 223, 85 wash drawings. London, 1948 (Jonathan Cape(15/-).—Dr. Swinton has admirably succeeded in his attempt to give a "simple account of the long history of living things." *The Corridor of Life* not only records the history of the inhabitants of this globe from the first beginnings of life to the present time, but also traces the main development of ideas concerning the origin of the earth and life upon it and the evolutionary principles which paleontology demonstrates.

The book commences with an adequate geologic introduction. Perhaps there is too much emphasis on methods of studying the earth's interior. Presentation of the concepts of geologic time and of modern ideas of the early history of the earth is excellent. Treatment of the principal invertebrates is more descriptive than evolutionary. Following a chapter in which vertebrates are contrasted with invertebrates and their origin discussed, somewhat over half of the book is devoted to their geologic history. Beginning with the early fishes, the trend of development is followed onto land with the emergence of the amphibians. Four chapters are devoted to reptiles (the author's special field), and the manifold adaptations of vertebrates to land, water, and air are illustrated from this group. Important mammalian types are summarized, and the evolution of horses and elephants—the two best examples—is considered in greater detail in the final chapter on "Evolution and Extinction." Here Dr. Swinton seems to favor teleological factors in evolution and to accept Broom's dictum that evolution is finished. However, in the final paragraphs, he turns this viewpoint into a strong argument for continued improvement of the human race through rational action.

The chief deficiency of this text is the omission of any discussion of fossil man and apes—a subject of great interest to the type of reader to whom such a book is addressed. There are minor inaccuracies, such as the statement that mastodons became extinct in the Pliocene and the illustration of the evolution of the horse, which fails to agree with the correct account in the text. The reconstructions by Miss Pinner are technically accurate but unfortunately tend toward the grotesque.

In spite of occasional looseness of terminology, *The Corridor of Life* is a good introduction to the biological side of earth history. The interested layman will find no difficulty in following this story, for technical terms are carefully explained where used. The book is particularly suitable for collateral reading in an historical geology course. Except for lack of detailed illustrations of the osteology of various animals discussed, it would be quite suitable as introductory paleontology text. Dr. Swinton is to be congratulated on preparing such a readable account of life in the past. JOSEPH T. GREGORY

Vibration and Sound; by PHILIP M. MORSE. Second Edition. Pp. xix, 468; 89 figs. New York, 1948 (McGraw-Hill Book Co., $5.50).—The second edition of this successful work on acoustical theory contains a very appreciable extension of the material presented in the previous edition. A larger page area and more than one hundred additional pages effect an increase of reading material of at least one-third. The most conspicuous—and welcome—feature of the new edition is the inclusion of the theory of transient acoustical phenomena. This analysis is set forth with the aid of operational calculas, providing an excellent illustration of the utility of this branch of mathematics. The first chapter carries a brief introduction to such constructs as contour integrals and Fourier transforms, and at the end of each subsequent chapter they are applied to the subject under discussion; the material is arranged, however, to allow their omission without loss of continuity in the book. As in the first edition, the emphasis throughout is on *methods* of acoustical analysis, and the author frequently stops to review and sum up the techniques he employs.

The chapter headings remain the same in the new edition but several new details have been added, besides the treatment of transients. Thus, piezo-electric crystals, the motion of a piano string, some theory of reed and wood-wind instruments, the absorption of sound by porous walls, and the general problem of transmission through ducts are a few subjects that add to the value of the second edition. The author has clarified and emphasized the role of the concepts of mechanical and specific acoustical impedance of various types of acoustical circuit elements. The transmission of sound inside cylinders is analyzed as a wave-guide problem, a procedure which suggests the feasibility of wider application of this method to the problems in the conduction and radiation of ultrasonic frequencies. The electrical analogue which has been so successfully exploited in acoustics gains new and greater value in the high frequency range.

On the debit side of the ledger there are very few items. The derivation of the equations of a sound wave seem awkward to this reviewer, and could profit from the introduction of the concept of velocity potential (the term is nowhere mentioned). One could also wish for some references to the literature of theoretical and experimental acoustics. A few typographical errors result from the bodily transfer of sections from the first to the second edition, but these are seldom serious enough to be confusing. Without doubt, Professor Morse has made a valuable contribution to the theory of sound. This book is a worthy shelf companion to Rayleigh's "Theory of Sound."

RICHARD F. HUMPHREYS

PUBLICATIONS RECENTLY RECEIVED

Arkansas Geological Survey. Bulletins as follows: 12 Molybdenum in Magnet Cove, Arkansas; by D. F. Holbrook. 13. Titanium in Southern Howard County, Arkansas; by D. F. Holbrook. Little Rock, 1948.

Theory of Equations; by J. V. Uspensky. New York, 1948 (The McGraw-Hill Book Co., $4.50).

Illinois Geological Survey. Bulletin No. 58. Oil and Gas Development in Illinois in 1947; by A. H. Bell, M. Aime, and V. Kline. Circular No. 146. Agstone used in Illinois in 1947; by W. H. Voskuil and E. M. King. Urbana, 1948.

New Mexico Bureau of Mines and Mineral Resources. Circular 18. Geology and Ore Deposits of Red River and Twining Districts Taos County, New Mexico. A preliminary Report; by C. F. Park Jr., and P. F. McKinley. Socorro, 1948.

Electromagnetic Waves and Light. An Introductory Physical Discussion. First Part; by C. F. Meyer. Ann Arbor, Michigan, 1948.

Crop Management and Soil-Conservation; by J. F. Cox and L. E. Jackson. 2nd Ed. New York, 1948 (John Wiley & Sons, $3.80).

Fundamentals of Physical Science. An introduction to the Physical Sciences; by K. B. Krauskopf. 2nd Ed.. New York, 1948 (The McGraw-Hill Book Co., $4.50).

Meson Theory of Nuclear Forces; by Wolfgang Pauli. 2nd Ed. New York, 1948 (Interscience Publishers, Inc., $2.00).

Organic Reactions. Vol IV; by Roger Adams—Editor-in-Chief. New York, 1948 (John Wiley & Sons, $6.00).

Rhenium; by J. G. F. Druce. Cambridge, England, 1948 (Cambridge University Press), New York, 1948 (The Macmillan Co., $2.50).

Physico Chemical Experiments; by R. Livingston. Revised Ed. New York, 1948 (The Macmillan Co., $3.50).

Crystal Structures; by R. W. G. Wycoff. First Section. Pp. 1-493, New York, 1948 (Interscience Publishers, Inc., $8.00).

Mathematical Biophysics; by N. Rashevsky. Revised Ed. Chicago, Illinois, 1948 (The University of Chicago Press, $7.50).

Kansas Geological Survey. Bulletin 73. Ground-Water Resources of Republic County and Northern Cloud County, Kansas; by V. C. Fishel. Lawrence, 1948.

American Journal of Science

APRIL 1949

METASOMATIC GRANITIZATION OF BATHOLITHIC DIMENSIONS[1]

PETER MISCH

PART I

ABSTRACT. In the Nanga Parbat area (Northwest Himalayas) a thick group of presumably pre-Cambrian argillites with thin calcareous and basic bands is overlaid by Cretaceous-Eocene volcanics, and both are bordered by extensive Eocene norite masses. These rocks have undergone regional metamorphism during Early Tertiary orogeny. Synkinematic metamorphism of the argillites progresses from slates and phyllites through micaschists, biotite-paragneisses, kyanite-schists and -paragneisses to sillimanite-paragneisses. The metamorphic zones in the argillites are matched by those in the calcareous bands.

Synkinematic granitization of the potash-predominant variety begins within the kyanite zone with microperthite-porphyroblast-schists passing to potash-rich augengneisses of granitic appearance which contain relict kyanite in their groundmass. Lit-par-lit replacement—as against mechanical injection—along active foliation planes makes banded gneisses. In the interior of the Nanga Parbat massif, granitization becomes more complete, and coarse-grained granitic gneiss forms. Here only small quantities of argillite altered to sillimanite-schist have escaped transformation into granitic rock. The calcareous bands are not granitized, and are preserved as conformable intercalations in the gneissose granite which has replaced the argillites. Metamorphism and granitization progress both along and across the strike of the altered sediments. There is complete structural continuity from the weakly metamorphosed sedimentary areas to the intensely granitized region. Also the type of tectonic deformation is the same throughout these areas, and is that characteristic of the deformation of solid rocks during orogenic metamorphism. The granitized body as mapped measures 25 by 60 miles.

The chemical nature of the granitizing agent is discussed. The metamorphic isograds (approximately vertical in this region) are shown to be independent of depth, and a function of differential introduction of heat from below. It is stressed that degree of metamorphism and granitization are systematically linked, and convection of heat by the granitizing solutions is held responsible for high grade regional metamorphisms in the upper part of the crust.

INTRODUCTION

THE idea that granitic rocks may form by metamorphic processes, is not new. It dates back to the French school of the last century, and to the subsequent classical contributions by J. J. Sederholm (1907, 1923, 1926, 1934). However, only during the last dozen years or so has the problem of granitization received ever-increasing general attention. There is no need for giving a historical review here, since several recent publications have done so, among which the Geological Society's granite memoir is outstanding. It is significant that various geologists, working in widely separated parts of the world, have arrived at more or less similar conclusions, quite independently of each other, and often without previous knowledge of published reports of earlier students of the problem. It appears that geology has at present reached a stage in which there is inherent a tendency to re-examine fundamental conceptions and theories sacred by tradition, and to substitute new ideas where critical evaluation of the ever-increasing body of observed facts so demands.

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[1] Read during the April 1948 annual meeting of the Cordilleran Section of the Geological Society of America at Pasadena.

This is especially true of the theory of the origin of granitic rocks—one of the basic problems in geology, considering the predominance of such rocks in the crust, and the role they play in the tectonic history of mobile belts. That granitic rocks can form by metamorphic transformation of pre-existing solid rocks on a comparatively small scale is today accepted by the great majority of geologists, since this process has been conclusively demonstrated by a number of authors, usually with an emphasis on small-scale features. Most geologists seem to feel that such processes are, as a whole, subordinate, and probably represent contact phenomena occurring in the marginal zones of igneous granitic batholiths. However, some of the workers in this field believe that metamorphic granitization is a process of much more fundamental and general significance. To prove this point, it has to be demonstrated that large masses of granitic rocks have as a whole, and not only marginally, formed by metamorphic transformation of solid rocks. The present paper attempts to contribute to this problem by describing some areas in which large masses of granitic rocks can be shown to have a metamorphic origin throughout.

The material presented was collected in the high mountains of Asia, but this is merely accidental; it appears to me that the areas described are not in any way extraordinary, but typical of what is found all over the world. The petrogenetic aspect of the Himalayan gneisses described below is identical with that of many gneisses ranging from pre-Cambrian to Mesozoic ages which I have seen in this country, in China and in Europe, just as the "small granodioritic blocks formed by additive metamorphism" described by G. E. Goodspeed (1937) from Northeast Oregon are duplicated in the Okanogan country in Eastern Washington, in the Indus Valley in the Northwest Himalayas, or near the Yangtse Valley in Northwest Yunnan.

The material presented below is arranged according to whether granitization was synkinematic, i. e. took place under conditions of active orogenic deformation, or whether it was postkinematic, i.e. took place under static conditions when orogeny had ceased. Then some cases are briefly described which might serve to illustrate the relationships between syn- and postkinematic granitization. At the end of each part, it is attempted to draw certain general conclusions, presuming that the material presented is typical.

FIRST PART

SYNKINEMATIC GRANITIZATION IN NANGA PARBAT AREA NORTHWEST HIMALAYAS

The Nanga Parbat area[2] in the crystalline inner zone of the Northwest Himalayas offers an unusually clear example of synkinematic granitization on a batholithic scale. Passages from metamorphic sediments of lowest to such of highest grades, and from chemically unaltered to intensely granitized rocks, are magnificently exposed in continuous sections for dozens of miles and over a vertical range of 23,000 feet. No structures or processes other or later than the main process of progressive alteration are superimposed, so that this process is revealed in grand simplicity.

Nanga Parbat Mountain rises north of the Great Himalayan Range, close to the upper Indus. The Nanga Parbat area shows a bending to north of the more westerly main trend of

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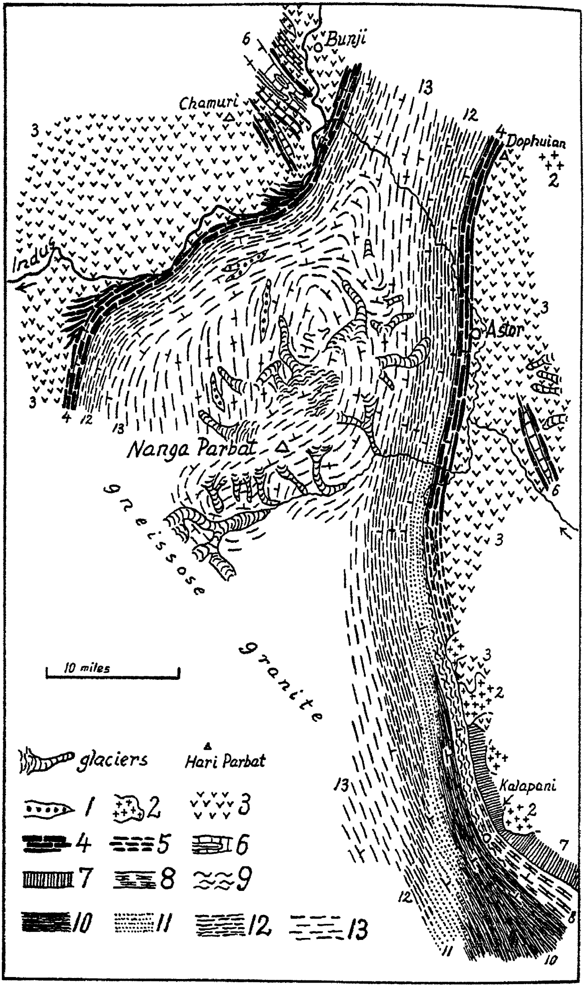
[2] The writer did four months of field work in this area during Willy Merkl's 1934 Nanga Parbat expedition. The petrographic study has occupied a number of years since. A report on field observations has been presented (Misch, 1935) but, apart from a short abstract (Misch, 1936a), publication of the petrographic results has been delayed by emigration to China and the outbreak of the Sino-Japanese war, and the ensuing lack of adequate facilities in the interior of China.


Bunji
Chamuri
Indus
Dophuian
Astor
Nanga Parbat
gneissose granite
Kalapani
10 miles
glaciers
Hari Parbat
1
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the Himalayas. This area forms a part of the eastern limb of the great syntaxial loop of the Northwest Himalayas which D. N. Wadia (1931) has described in detail in the sedimentary zones farther south.

(I) MAIN ROCK UNITS OF NANGA PARBAT AREA

The least metamorphosed bedded rocks are found in the southeast of the area shown in text figure 1, and farther south. Here occurs a *group of black phyllites and slates* with subordinate thin limestone bands, and occasional thin layers of quartzite and altered greenstone. The original thickness is very great and tectonic accumulation by intense folding is added. The resulting thickness exceeds six miles in the southeastern part of the region shown in figure 1, and south of this region the phyllite-slate group forms a still much thicker section. In the southeastern part of the region, this group is succeeded on the east by variegated tuffaceous and calcareous slates which in their upper part are interbedded with tuff breccias. Still higher in the section, there appear basic and also more acidic lava flows which in their lower part are intercalated with the tuffaceous rocks. This *volcanic formation* has been shown by D. N. Wadia (1937) to be *Cretaceous-Eocene*, on the basis of fossils discovered further southeast. Wadia

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Text Fig. 1. Geologic sketch map of Nanga Parbat region, Northwest Himalayas. After P. Misch (1935), fig. 8.

1: tourmaline-granite; postorogenic
2: quartz-diorite and granodiorite; post-norite
3: norite, hypersthene-diorite, etc.; widely altered to metanorite and metadiorite; Eocene
4: mesozonal amphibolite } synorogenic-metamorphic norite etc.
5: epizonal greenschist }
6: marble and schist zones in norite, metanorite, etc.
7: basaltic and more acidic lavas } Cretaceous-Eocene volcanic formation
8: variegated tuff breccias and tuffaceous-calcareous argillites }
9: tuffaceous argillaceous rocks (8) epizonally metamorphic }
10: Salkhala group; dark slates and phyllites with subordinate thin limestone and greenstone bands }
11: non-granitized schists; chiefly mesozonally metamorphic Salkhalas }
12: mesozonal migmatitic gneisses (chiefly augengneisses and banded gneisses); incompletely granitized Salkhalas } metamorphism and granitization synorogenic
13: migmatitic granitic gneiss; granitized Salkhalas }
(11)-(13) contain intercalations of schists, marbles, and amphibolites; mesozonal in (11) and (12), katazonal in (13) }

(1932) assigns the underlying black slate group to the *pre-Cambrian Salkhala series* of the Western Himalayas. According to this author, the Salkhalas are everywhere separated by an angular unconformity from the Late Algonkian Dogra slate and all younger rocks. In the area here discussed, no angular relationship is visible in the field between the black slate group and the volcanic formation, their directions of foliation being parallel. In the tuffaceous rocks the foliation is essentially parallel to the bedding. In the black slates and phyllites, no bedding is visible in the field. Under the microscope, some of these rocks show intensely folded fine-bedding traversed by the foliation; isoclinal folds have their limbs parallel to the foliation and become drawn-out and finally obliterated by slip along the foliation planes. Any angular unconformity which may originally have existed is thus masked by the foliation.[3] The foliation is later than the volcanic formation and has originated during the late or post-Eocene main Himalayan orogeny.[4]

The black slates and phyllites and their limestone and greenstone bands pass gradually into *mesozonal crystalline schists* toward the west and north, that is, both across and along the strike. After progressive metamorphism has reached a mineral facies corresponding to the lower part of the mesozone, potash-feldspar porphyroblasts appear in the dark-colored schists[5] (chiefly biotite-paragneisses[6]) of argillaceous

[3] According to Wadia, such intense minor folding is characteristic of the Salkhalas though later structures may be superimposed.

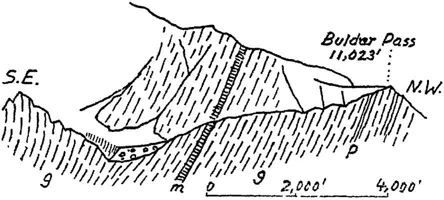
[4] The Himalayan folding is often called "Late Tertiary." However, the folding of the geosynclinal main range is Early Tertiary. Only the foldings of the successive foredeeps, marginal thrusting of the main range over the fore-deep rocks, and some local disturbances within the main range are Late Cenozoic (cf. Misch, 1936b, with references).

[5] The term *schist*, without a prefix, is here used to indicate any schistose rock derived from an argillaceous sediment, and having reached a higher metamorphic grade than phyllite. "Schist" thus includes micaschist, paragneiss and related rocks. "Crystalline schist" is a comprehensive term including all foliated metamorphic rocks. Schistose rocks differing in their composition from schists of argillaceous derivation, are indicated by a qualifying prefix or adjective, e.g. "chlorite-schist," "greenschist" (epidote-albite-chlorite-schist), "actinolite-schist," "calcareous schist," "quartzite-schist," etc.

[6] The term *paragneiss* is here used for crystalline schists having as main constituents quartz, mica and feldspar, and derived from sediment. Most paragneisses are fine-grained, rich in biotite and dark-colored. They may be thinly foliated, but are mostly somewhat more compact. They differ from micaschists in having feldspar as one of their main constituents, and mostly in being less thinly foliated.

derivation which thus grade into *mesozonal migmatitic augen-gneisses and banded gneisses*, both across and along the strike. The mesozonal rocks form a continuous belt on the eastern side of the Nanga Parbat gneiss massif, and a similar belt occurs in the northwest on the opposite side of the massif.

In the mesozonal belts, granitization remains incomplete, and material of purely sedimentary derivation is still recognizable everywhere. Toward the *interior of the gneiss massif*, granitization reaches a much higher degree. Most of the argillite-derived fine-grained dark-colored paragneiss becomes more coarsely recrystallized and feldspar-enriched, and thus loses its distinctive character. Here the mesozonal migmatitic gneisses grade into *coarse-grained light-colored granitic gneisses or gneissose granites*. These rocks form the bulk of the interior of the gneiss massif. Layers of dark-colored argillite-derived katazonal paragneiss, and of banded gneiss containing clearly discernible paragneiss laminae, are quantitatively very subordinate, but are intercalated throughout the interior of the massif. The granitic gneiss also contains layers of kata-amphibolites and of marbles with katazonal lime-silicates. These rocks may be associated with some paragneiss, but also occur directly intercalated in granitic gneiss. Particularly striking are a few places where pure white marble forms regular bands of small thickness but great longitudinal extent immediately in the coarse-grained gneissose granite (cf. text figure 2).



Text Fig. 2. Section and view south of Buldar Pass, northeast of Nanga Parbat.

g: coarse-grained granitic gneiss
m: white marble
p: fine-grained dark-colored biotite-schist (paragneiss)

Field observations suggest, and microscopic structures confirm, that the progressive regional metamorphism is *synkinematic*. The same holds true of the process of granitization though in the interior of the gneiss massif there are indications that crystallization in general and granitization in particular continued after deformation ceased.

In addition to the black slate group (Wadia's Salkhalas), the *Cretaceous-Eocene volcanic formation* occurring in the southeast of the area, *has undergone regional metamorphism.* In the extreme southeast of the area, alteration does not go beyond the slaty stage, and disappears on the east in the upper members of the volcanic formation. Toward the north, however, the metamorphism of the volcanic formation increases along the strike. Chlorite-schists, calc-phyllites, intensely laminated tuff-breccias with chloritic matrix, etc. represent the epizone. Farther north, hornblende-garbenschiefer and calc-micaschists, often rich in epodite-clinozoisite minerals, show the facies of the uppermost mesozone. Still farther north, some normal mesozonal amphibolites probably prolong this stratigraphic unit.

Another group of rocks in the Nanga Parbat area consists of very extensive masses of *norite and hypersthene-diorite* with local dunite. These rocks border the gneiss massif both on the east and northwest. The eastern norite area is on the south in contact with the predominantly basic lavas of the Cretaceous-Eocene volcanic formation. The norite is contemporaneous with or slightly younger than the lavas. That both are genetically related has already been pointed out by D. N. Wadia (1932).

Most of the noritic rocks have been more or less metamorphosed. There are all passages to statically recrystallized metanorites and metadiorites with crystalloblastic hornblende, epidote, etc. As the borders of the gneiss massif are approached, static recrystallization gradually gives way to kinematic metamorphism,[7] and *mesozonal amphibolites* are formed. The mineral association remains the same, except for the appearance of garnet in the deformed rocks. At the borders of the gneiss massif, the norite-amphibolites are

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[7] The noritic series, its relation to large areas of included metamorphic Salkhalas where alternations and gradational contacts are observed, the later metamorphism of the noritic rocks, and the relations between their static and kinematic types of metamorphism, will be dealt with in another paper.

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TABLE I. MINERAL RANGES* IN METAMORPHIC SEDIMENTS OF NANGA PARBAT REGION, NORTHWEST HIMALAYAS

temperature zones	EPIZONE "highest" "lowest"	MESOZONE "upper"	"lower"	KATAZONE
ARGILLITES (in ARGILLACEOUS QUARTZITES: same minerals, with larger proportion of quartz)	SERICITE SERICITE or MUSCOVITE always present	MUSCOVITE usually much	often less, sometimes absent	MUSCOVITE less or absent ("conditionally stable")
	CHLORITE frequently present	BIOTITE nearly always present		BIOTITE always present
	BIOTITE sometimes pres. (mostly sericitiform)		KYANITE frequently present	SILLIMANITE often present
	Mn-rich garnet sometimes present	ALMANDITE commonly present		ALMANDITE often present
		POTASH FELDSPAR frequently present		ORTHOCLASE nearly always present
	ALBITE ALBITE or ALB-OLIGOCL. often present	OLIGOCLASE widely present	nearly always present	OLIGOCLASE nearly always present (more calcic plagioclase only in calcareous argillites)
	QUARTZ always present			QUARTZ present except in rare cases
CALCAREOUS ARGILLITES and ARGILLACEOUS LIMESTONES	SERICITE SER. or MUSC. always present	MUSCOVITE frequently present	less or absent	DIOPSIDE always present (unless rock rather low in Ca + Mg in which case hornblende forms; diopside may also be absent in rocks extremely high in Ca + Mg, e.g. in forsterite - bearing marbles)
	CHLORITE frequently present	BIOTITE mostly present	nearly always present	
	BIOTITE often present (often sericitiform)	POTASH FELDSPAR often present		accessory ORTHOCLASE present unless rock is extremely low in K_2O
	Mn-rich garnet sometimes present	ALMANDITE commonly present		GROSSULARITE mostly present in hottest katazone no grossularite
	ALBITE ALB - OLIGOCL or ALBITE often present	OLIGOCLASE widely present	nearly always present	BASIC PLAGIOCLASE (mostly bytownite) always present except in some rocks extremely high in Ca + Mg
	EPIDOTE-CLINOZOISITE frequently present	EPIDOTE-CLINOZOISITE & ZOISITE frequently present	ZOISITE commonly present EPIDOTE often present CLINOZ. less common	WOLLASTONITE frequently present
	CALCITE (DOLOMITE) always present	CALCITE always present		CALCITE present if $Ca+Mg > Si$, or if reaction incomplete
	QUARTZ nearly always present			QUARTZ present if $Ca+Mg < Si$, or if reaction incomplete
			SCAPOLITE often present	HORNBLENDE present if rock rather low in Ca + Mg ("conditionally stable")
	in Mg-rich rocks (dol shales) ACTINOLITIC HORNBLENDE sometimes present	in Mg-rich rocks (dolomitic shales) HORNBLENDE frequently present sometimes to exclusion of mica		in high Ca + Mg - excess rocks (impure marbles) FORSTERITE, HUMITE, SPINEL, TREMOL., PHLOG.
types of synkinem. granitized rocks			FELDSPAR PORPHYROBLAST SCHIST AUGENGNEISS BANDED GNEISS	GRANITIC BIOTITE GNEISS or GNEISSOSE GRANITE

* rutile (in argillites), sphene (in impure calcareous rocks), iron oxides, zircon, apatite, tourmaline, orthite, etc. are omitted

intimately intercalated and interfolded with the paragneisses and migmatitic gneisses. Both rock units, e. g. the norite bodies and the gneiss massif, have here participated in the same process of tectonic deformation and of crystallization accompanied by some granitization. This process of synkinematic metamorphism took place during the Early Tertiary main Himalayan orogeny, not a very long time after the original norite bodies had formed. At contacts with granitized bands, the amphibolites tend to be locally biotized.

Whereas the metamorphism of the norites is generally mesozonal, it becomes epizonal (chlorite - epidote - albite - schists, etc.) along the western border of the eastern norite body, where this border approaches the epizonal southeastern portion of the Nanga Parbat area.

Directionless granitic rocks occur in several parts of the area, but remain subordinate. Small irregular patches and pockets of quartz-dioritic and granodioritic rocks showing gradational contacts and porphyroblastic textures are common in directionless or weakly schistose metanorites. Some larger bodies of granodiorite occur in the noritic rocks and the lavas of the volcanic series. Within the gneiss massif, dikes and larger bodies of postorogenic white tourmaline-granite occur at several places, but the largest attain a size of only a few miles.

The *boundaries* between the metamorphic rock series described are all steep. The *metamorphic isograds* are steeply inclined planes in the whole area. For the most part, they are nearly perpendicular, the dip being either normal or slightly overturned.

(II) PROGRESSIVE METAMORPHISM OF SEDIMENTS IN NANGA PARBAT AREA

Argillite series.—This is the problem confronting us: do the granitic gneisses which form the bulk of the Nanga Parbat massif, owe their origin to intrusion of granitic magma, or to metasomatic[8] granitization? I shall try to give evidence for the latter process, e. g. for a formation of the granitic gneisses by synkinematic metamorphic transformation and replacement of sediments, concurrent with certain chemical changes which chiefly consisted of the addition of alkali. This problem cannot be approached without a previous discussion of the progressive metamorphism the non-granitized sediments have undergone. Moreover, granitization should not be visualized as an isolated phenomenon, but in its integral relationship to higher-grade regional metamorphism.

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[8] The term "metasomatic" is here used in its widest sense, i.e. to indicate any alterations of rocks during which chemical changes are brought about by an introduction of new substances, which process may or may not involve a simultaneous removal of other substances.

Most of the sediments in the Nanga Parbat area are argillites belonging to the Salkhala group, and the sedimentary base of the granitic gneisses almost entirely consists of these rocks. Therefore, the series of progressively metamorphosed argillites is the one with which we are chiefly concerned. It shows, in a continuous succession observed both across and along the strike, *a number of "mineral facies"* (cf. P. Eskola, 1920) which can be readily assigned to the three classical *"depth" zones.* The writer does not know another area in which the zones of regional metamorphism are displayed with equal completeness and continuity and in equally magnificent exposures. The *factor controlling the spatial distribution of mineral facies and zones* in this area can be shown to be exclusively *temperature.* Confining pressure was approximately the same in places greatly differing in their degree of metamorphism. For the metamorphic zones succeed each other horizontally, with approximately vertical boundaries; in fact, most of the katazonal rocks forming the interior of the Nanga Parbat massif are exposed at higher elevations than most of the meso- and epizonal rocks of the marginal areas. Stress action was also of the same intensity in the different metamorphic zones as is shown by the structures of the rocks although, as is to be expected, crystallization has healed cataclastic mineral structures more completely in the hotter inner portions of the massif than in the marginal areas.

The series of metamorphosed argillites begins with black carbonaceous slates and ends with high-grade sillimanite-bearing rocks. The upper epizone is represented by carbonaceous *phyllitic slates* composed of sericite and quartz, often with additional chlorite. They pass into less finely textured normal *epizonal phyllites* (cf. pl. 1, fig. 1) composed of the same minerals, often with the addition of albite; the albite tends to form small porphyroblasts which frequently show helicitic structures and sometimes bear signs of having been rolled. In the albite-free low-grade rocks, some sodium is probably camouflaged in sericite (admixture of paragonite). In the *lower part of the epizone,* small porphyroblasts of supposedly manganese-rich garnet appear, still in association with chlorite.[9] At the *base of the epizone,* the phyllites become more highly crystalline and approach micaschists. The essential minerals are muscovite and quartz. Chlorite still survives as a contemporaneous (not retrogressive) constituent and assigns these rocks to the epizone although members of the biotite series occur in some of them as a predominantly fine-grained constituent resembling sericite in texture. The albite may pass into oligoclase-albite. Garnet is common. In the various epizonal Salkhala argillites described, calcite, epidote and clinozoisite are only locally found, since these argillites are usually poor in calcium. On the other hand, these minerals are common in calcareous slates and phyllites of the Cretaceous-Eocene volcanic formation. Otherwise, those rocks show the same mineral assemblages as the Salkhala argillites.

As the mesozone is entered, contemporaneous chlorite disappears. The *upper part of the mesozone* is characterized by schists composed of well-formed biotite and muscovite, and of quartz (cf. pl. 1, fig. 2). Occasionally muscovite becomes subordinate to biotite or even disappears, obviously owing to a relatively high content in iron and magnesium, and to the lack of a pronounced aluminum excess. Almandite is a common accessory and tends to form porphyroblasts; "snow-ball garnets" are common and indicate, the same as other structural features, that crystalloblastic growth and mechanical deformation were simultaneous. There is often some additional oligoclase (usually on the sodic side, owing to a low calcium content of the sediment). If the amount of oligoclase is increased, the micaschists pass into paragneisses (cf. pl. 2, fig. 1), according to the definition of the term given above. Potash-feldspar appears within the mesozone, but not at its upper boundary. None of the feldspars show porphyroblastic tendencies in these micaschists and paragneisses. The amount of feldspar rarely reaches that typical of the granitic gneisses.

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[9] The writer cannot follow the widely used classification "chlorite zone," "biotite zone," "garnet zone," since in the rocks studied by him the ranges of stability of these minerals overlap.

The *lower mesozone* is characterized by the appearance of kyanite. This mineral does not occur in all argillites metamorphosed in this zone, but only in those which have a pronounced aluminum excess, that is, in rocks derived from rather pure clays. In these rocks, kyanite often becomes one of the main constituents (cf. pl. 2, fig. 2). The other minerals are the same as in the upper part of the mesozone. However, there seems to be a tendency to somewhat reduce the amount of mica and especially of muscovite, and accordingly to increase the amount of feldspar. The crystallization of free aluminum-silicate accounts for this, most of the aluminum excess having been fixed in muscovite before the temperature had become high enough to permit formation of kyanite.[10] Accordingly, most of the kyanite rocks are paragneisses rather than micaschists, and potash-feldspar is rarely absent. Occasionally muscovite is reduced to the role of an accessory or is even absent. However, biotite-free varieties exceptionally occur; they are rocks very low in magnesium and with a moderate iron content which was fixed in the almandite present in almost all the kyanite rocks. In exceptional cases, kyanite rocks are so low in feldspar that they must be classified as micaschists, the limited amount of potash present having all been used up by the mica, and the quantity of sodium having only been sufficient to make some accessory oligoclase. Apart from kyanite rocks, the lower mesozone contains "ordinary" micaschists and paragneisses hardly distinguishable from similar rocks of the upper mesozone (cf. pl. 2, fig. 1).

In the *katazone*, *sillimanite* takes the place of kyanite (cf. pl. 3, fig. 1). All sillimanite-rich rocks collected were found to contain a considerable amount of potash feldspar usually represented by sanidine, and generally a considerably lower amount of oligoclase. At the same time, biotite considerably exceeds muscovite, and this latter may even be absent; high temperature which led to the crystallization of almost all of the aluminum excess as sillimanite, is the obvious explanation.

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[10] It is of course realized that the amount of water available is another factor controlling the relative amounts of mica and feldspar. However, this factor can here be neglected since there are indications that there was no shortage of water in any of the metamorphic zones in this area.

Like the kyanite-rich rocks, those rich in sillimanite are derived from rather pure clays. Some of the sillimanite-rich rocks are practically free of quartz. Along with the high aluminum excess, this underlines the pure nature of the original clays from which this particular variety of rock was derived.

As a whole, non-granitized metamorphosed argillites occur only in very subordinate relict bands in the katazonal interior of the Nanga Parbat massif. Most of these bands do not contain sillimanite, but are "ordinary" paragneisses. Their mineral composition does not differ from that of the "ordinary" paragneisses of the mesozone, except for frequent absence of muscovite.[11] As in the mesozone, the oligoclase is usually on the sodic side. This is due to a primary low calcium content. Where katazonal paragneisses are in contact with lime-silicate bands, their plagioclase becomes at once more basic. The low calcium content of the meso- and katazonal paragneisses is also demonstrated by the presence of rutile whereas in metamorphosed calcareous shales of both zones the titanium has combined with calcium to form sphene. Like the mesozonally metamorphosed argillites, those of the katazone lack porphyroblastic development in their feldspars. In all these rocks, the carbonaceous matter characteristic of the Salkhala slates is still widely present in the form of finely scattered graphite flakes.

The textures of the meso- and katazonally metamorphosed argillites are fine-grained. Most of these rocks are dark-colored, owing to their high content in finely divided biotite. The appearance of the metamorphosed argillites is very different from that of the granitized rocks which show light colors, coarser textures, and porphyroblastic feldspars.

Lime-silicate series.—The argillite series is perfectly matched by the series of progressively metamorphosed impure calcareous rocks occurring as subordinate thin bands in the Salkhalas and, in the eastern portion of the area, in the lower part of the Cretaceous-Eocene volcanic formation. Considering the small amount of calcareous rocks in the Salkhalas, and in view of the fact that they have usually escaped granitiza-

[11] The writer cannot agree with the opinion held by many authors that muscovite is a stranger to the katazone. He thinks that muscovite, like hornblende, is "*unconditionally stable*" in the mesozone, and is "*conditionally stable*" in the katazone (meaning, stable under certain limited chemical conditions). This subject will be treated elsewhere.

tion, their metamorphism is only briefly discussed in the present paper.[12]

Upper-epizonal calcareous slates and *epizonal calc-phyllites* contain the same minerals as the non-calcareous argillites, with additional calcite (dolomite), epidote and clinozoisite. Also here, fine-grained members of the biotite series appear before chlorite becomes unstable. Calc-micaschists with either biotite or muscovite or both, are frequently rich in epidote or zoisite; they range close to the boundary of epi- and mesozone, and seem to belong to the highest mesozone. To the same mineral facies belong *calc-micaschists with hornblende porphyroblasts*; part of them are developed as *garbenschiefer.* These rocks are derived from dolomitic shales whereas the epidote-zoisite-rich micaschists represent calcareous shales. Varieties rich in magnesium and very poor in potash become para-amphibolites free of mica. In some of these rocks, the green hornblende is still associated with contemporaneous chlorite which assigns this variety to the lowest epizone.

The majority of the rocks just described belong to the Cretaceous-Eocene volcanic formation although bands of calcareous phyllites and micaschists are also found in the Salkhalas, often together with marble layers. In the *upper mesozone*, such bands are characteristically developed as *calcareous zoisite-bearing biotite-muscovite-quartz-schists*, they usually contain oligoclase and often almandite. Within the *lower mesozone*—corresponding to the kyanite-schist facies in the pure argillites—scapolite appears in the impure calcareous bands intercalated in the Salkhalas. The rocks are *scapolite-paragneisses* and, more rarely, scapolite-micaschists. Their minerals are· scapolite, zoisite, biotite, muscovite (often absent), oligoclase, potash-feldspar, calcite, quartz, almandite, and sphene. The latter mineral is common in impure calcareous rocks from epi- to katazone. 'Impure epi- and mesozonal marbles contain the same minerals as the schists described.

In the katazone, the association calcite + quartz becomes unstable in the impure calcareous bands in the Salkhalas. Now true lime-silicate minerals form. However, not infrequently the reaction remains incomplete, presumably owing to the CO_2 being unable to escape (cf. V. M. Goldschmidt, 1911). In the upper katazone, basic plagioclase and diopside appear and scapolite still survives from the lower mesozone (unless it is retrogressive). Examples are: hornblende-bytownite-granulite with quartz, calcite and garnet, forming a band in biotite-paragneiss; diopside-scapolite-marble and granulite, with basic plagioclase and some potash-feldspar, occurring in a similar way.

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[12] This subject will be more fully discussed in another paper.

With further increase in temperature, the *katazonal lime-silicate-granulite bands* reach their typical development and wollastonite is added as a characteristic constituent, representing the same mineral facies in which sillimanite forms in pure argillites. The wollastonite is associated with bytownite or bytownite-anorthite. This paragenesis is remarkable since, according to V. M. Goldschmidt's classical memoir on the contact metamorphism in the Oslo region (1911), this association could not be stable and the two above minerals ought to react to form grossularite. However, in the Nanga Parbat rocks wollastonite and bytownite are intergrown as contemporaneous and obviously stable partners. Since pressure would favor the reaction, and since the Nanga Parbat rocks have undoubtedly formed under considerably higher pressure than the Oslo and similar contact rocks, it must be concluded that in the interior of the Nanga Parbat massif *temperature was considerably higher than in the "inner contact zone,"* and was thus able to overcome the not inconsiderable pressure and prevent the grossularite reaction. This is confirmed by the presence, in part of the Nanga Parbat lime-silicate-granulites, of later grossularite replacing the paragenesis wollastonite + bytownite. The grossularite reaction began to take place when temperature, though still katazonal, had become somewhat lower.

The magnesium of the lime-silicate-granulite bands is fixed in diopside. Where such bands are in contact with ordinary paragneiss, a boundary zone rich in common green hornblende usually intervenes, demonstrating that hornblende, containing less Ca as well as Ca + Mg than diopside, is "conditionally stable" in the katazone (cf. above, foot-note no. 11). As the rock grades into purer argillite, biotite takes the place of hornblende. In the lime-silicate bands, biotite never occurs, and the potash is fixed in some accessory orthoclase. The lime-silicate-granulites contain either excess quartz or excess calcite, but sometimes both are present (incomplete reaction, cf. above). Bands of *lime-silicate-marble* often contain forsterite, instead of diopside (most economic use of limited silica available). If aluminum was available and there was a shortage of silica, spinel has formed in marble bands. Also, phlogopite is found in marbles of this zone.

The gradation, biotite - paragneiss, hornblende - granulite, lime-silicate-granulite, lime-silicate-marble, pure marble, is often observed within a few centimeters and even less, and close and regular interbanding is common. It is thus indicated that the primary differences in chemical composition which can be traced back to but slightly metamorphosed Salkhala sediments, have been perfectly preserved even under katazonal conditions, and that *practically no chemical migrations have taken place within the calcareous rocks*. This fact is remarkable, in view of the considerable chemical changes which have transformed most of the metamorphosed argillites into granitic gneisses (cf. below). The preservation of pure marble layers, not only next to paragneiss, but even within (migmatitic) granitic gneiss, illustrates the lack of chemical migrations in the calcareous bands.

The progressive metamorphism of argillaceous and impure calcareous rocks at Nanga Parbat is summarized in table 1. The *progressive metamorphism of basic rocks* matches that of the sediments. Basic rocks occur in the Cretaceous-Eocene volcanic formation, in the norite masses, and, as subordinate intercalations, in the Salkhalas and the gneisses formed from them. There is no need to describe in the present paper the various mineral facies displayed by the basic rocks. It may only be mentioned that in the sillimanite-gneiss and lime-silicate-granulite facies, there is no sign of an "eclogite facies" in the basic rocks, but they are developed as katazonal amphibolites; obviously, the eclogite facies is not a normal katazonal facies, but one restricted to extreme pressure, e. g. great depth.

(III) SYNKINEMATIC GRANITIZATION IN NANGA PARBAT MASSIF

(*a*) *Outer zones*

After metamorphism of the Salkhala argillites had reached a certain stage within the lower mesozone, their granitization began. Its intensity steadily increased toward the katazonal interior of the gneiss massif.

Both on the eastern and western side of the Nanga Parbat massif, granitization set in within the kyanite zone. First smaller and then large porphyroblasts of potash-feldspar began to form in the fine-grained schists (cf. pl. 3, fig. 2). As the number of porphyroblasts increases, the *feldspar-porphyroblast-schists* grade into *coarse-grained augengneisses* of granitic appearance (cf. pl. 4, fig. 1) and, if porphyroblasts are aligned in stringers and coalesce into bands, they grade into *banded gneisses* (cf. pl. 3, fig. 2; pl. 4, figs. 2-4). Both kinds of metamorphic migmatites are usually rich in almandite. There are all kinds of gradations, intermediate types, interlayering and interfingering of these variously granitized rocks, as well as of non-granitized metamorphic argillites.

(1) *Mesozonal migmatitic augengneisses.*—Both on the eastern and western side of the Nanga Parbat massif, there is a *continuous belt* thousands of feet wide in which massive coarse-grained augengneisses greatly predominate. These belts occupy the outermost parts of the massif of granitized Salkhalas. This augengneiss belt is absent only in the south-eastern part of the area shown in figure 1.

Although having an igneous appearance (pl. 4, fig. 1), these massive augengneisses reveal their metasomatic origin in the field, by the presence of all gradations between augengneisses and argillite-derived schists and in the interlayering of both kinds of rock (cf. Misch, 1935). Microscopic examination confirms the metasomatic-migmatitic nature of the augengneisses. Their *groundmass* is, in its mineral composition and texture, identical with the biotite-schists and paragneisses described above which are altered Salkhala argillites. In this matrix, the large *potash-feldspar porphyroblasts*, usually microperthite, grew, replaced and penetrated it, and enclosed relics of its mineral grains (pl. 5, fig. 1; pl. 6, fig. 2; pl. 7, fig. 1). The irregular and crenulated borders of the porphyroblasts finally approach crystal faces, as the porphyroblasts tend to become idioblastic (pl. 6, fig 2). Karlsbad twinning is common. The amount of porphyroblasts become very large, they exceed half of the volume of many rocks, and the original schist becomes restricted to relict stringers and winding bands between the large feldspars. Of particular interest is the presence of kyanite in the argillite-derived matrix of many of the augengneisses (pl. 5, fig. 1; pl. 7, fig. 1). Projections of growing feldspar porphyroblasts can be seen invading kyanite-rich paragneiss fabric, and absorbing it (pl. 7, fig. 1). Pure aluminum-silicate would hardly crystallize in a rock so rich in potash that more than half of it consists of potash-feldspar (not to mention the additional potash fixed in mica). In other words, the kyanite chemically represents an *unstable relic.* One chemical equilibrium was succeeded by a new and different one, and the difference was chiefly due to an introduction of potash.

The texture of the rock indicates that the potash-feldspar porphyroblasts have not been introduced as a whole, but are the product of a *reaction between the argillite-derived schist and newly introduced material* rich in potash. The introduced material was in a condition which enabled it to infiltrate and permeate the country rock, and to react with it. This means that the introduced material consisted of *hot solutions.*[13] The formation of the potash-feldspar porphyroblasts cannot be attributed to metamorphic differentiation within the country rock, both because the amount of potash in the augengneisses is higher than in the argillite-derived schists in the non-granitized areas, and because the matrix of the augengneisses possesses the original mineral composition of the argillite-derived schists and cannot therefore be a differentiate.

The *amount of material added* was not large (no exact figures can be given unless chemical analyses are made). A normal argillite is not too far from a granite in its chemical composition, and an addition of a moderate amount of alkali and usually some silica will transform it into a rock of granitic composition; whether silica is needed depends on the amount of quartzose silt admixed with clay minerals in the original argillites. In the mineral composition of these augengneisses, there is no indication of an introduction of sodium at this stage. However, there is a possibility that some sodium was introduced already at an earlier stage of metamorphism, namely in those mesozonal paragneisses, which contain a considerable amount of oligoclase. Since these rocks are fine-grained and lack porphyroblastic feldspars, there is no textural indication of an introduction of sodium, and it could be ascertained only by chemical analyses. Even if there has been some addition of sodium, these fine-grained and even-textured paragneisses could not possibly be described as "granitized."

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[13] Several authors have recently emphasized the importance of solid diffusion in chemical migrations during metamorphism Though it has been demonstrated that such a process is possible, the writer's observations lead him to believe that solid diffusion will, if at all, usually take place only on a very minor scale (as between adjacent minerals—albitization of basic plagioclase where invaded by newly formed hornblende, observed by the writer in metanorites), but can hardly be held responsible for major chemical transfers during regional metamorphism and granitization; the action of solutions is often shown by textural features connected with the intergranular film, and that water was present is indicated by the occurrence of hydrous silicates, as well as by general considerations (water content of sediments, especially of argillites, phyllites, micaschists, etc.).

Unless an equal amount of material has been removed in solution from the country rock—of which there is no indication in the rock successions in question—the *volume* must have been somewhat increased by granitization. Such a small increase in volume would not be easy to prove even in a rock granitized under static conditions, and is much more difficult to ascertain in these rocks which have been metamorphosed and granitized during mechanical deformation. The winding shape of bands of paragneiss material between the feldspar porphyroblasts might be considered as an indication of increase of volume in the areas replaced by porphyroblasts. That the augengneisses are *synkinematic* is shown by the elongation and alignment the porphyroblasts have acquired during their crystalloblastic growth, and by cataclastic deformation the porphyroblasts have suffered at some places where movement outlasted growth, whereas in other places all cataclastic structures have been healed by crystallization.

(2) *Mesozonal migmatic banded gneisses.*—Banded gneisses occur as intercalations within the belts of massive augengneiss, but have their main development on the inner side of these belts. Here they form *zones several miles wide* on both sides of the Nanga Parbat massif. In the southeast of the area where the augengneiss belt disappears, the zone of banded gneisses grades into the area of non-granitized argillite-derived schists. The banded gneisses consist of alternating layers of dark-colored biotite-schist and paragneiss, and of white or light-colored gneiss of the appearance of an acidic orthogneiss (pl. 3, fig. 2; pl. 4, figs. 2-4).

The *biotite-schist and paragneiss bands* are, in composition and texture, identical with the rocks occurring in the non-granitized areas of metamorphic Salkhala argillites. Kyanite frequently occurs in these bands (pl. 5, fig. 2; pl. 6, fig. 1). The *light-colored bands* (pl. 7, fig. 2; pl. 8) are for the most part less fine-grained, and are leucocratic and rich in feldspar, especially potash-feldspar (frequently microcline). Muscovite is generally present in addition to potash-feldspar, oligoclase (usually sodic), and quartz. Biotite occurs in much smaller amounts than in the dark-colored bands, and is often altogether absent. Almandite and tourmaline are common accessories both in the light- and dark-colored bands. Tourmaline is also widely developed in the non-granitized areas of metamorphic argillites, and therefore can hardly be considered as a "pneumatolytic" mineral, but may have been derived from the boron originally contained in the marine argillites.

Apart from white bands with little or no biotite, there are also *bands intermediate in composition* and color between the dark-colored schists and the white bands. All gradations and all kinds of interbanding occur (pl. 4, figs. 2 and 3). The *thickness of individual bands* of dark-colored schist and light-colored gneiss is widely variable, both ranging from a few millimeters to many feet. The *proportion* of both kinds of rock in the banded gneisses is likewise extremely variable. The series begins with metamorphic argillites containing only a few thin white bands usually accompanied by individual feldspar porphyroblasts (pl. 3, fig. 2), and it ends with white acidic gneisses in which only subordinate thin layers of schist have been preserved (pl. 8). Banded gneisses showing different proportions are intercalated in numerous repetitions.

There occur *all passages between banded gneisses, augengneisses, and feldspar-porphyroblast-schists* (pl. 3, fig. 2; pl. 4, fig. 4; pl. 5, fig. 2; pl. 6, fig. 1), and subordinate layers of typical augengneisses may be intercalated in the main zone of banded gneisses. Across the strike, light-colored bands frequently grade into schists with potash-feldspar porphyroblasts (pl. 4, fig. 4). Along the strike, light-colored bands frequently die out within the schist, or are continued by layers rich in potash-feldspar porphyroblasts (pl. 4, figs. 2 and 4). Looking at such layers from the other end, their porphyroblasts may be said to coalesce finally into white bands. Also within the light-colored bands, the feldspar frequently shows a *porphyroblastic development*, often discernible in the field

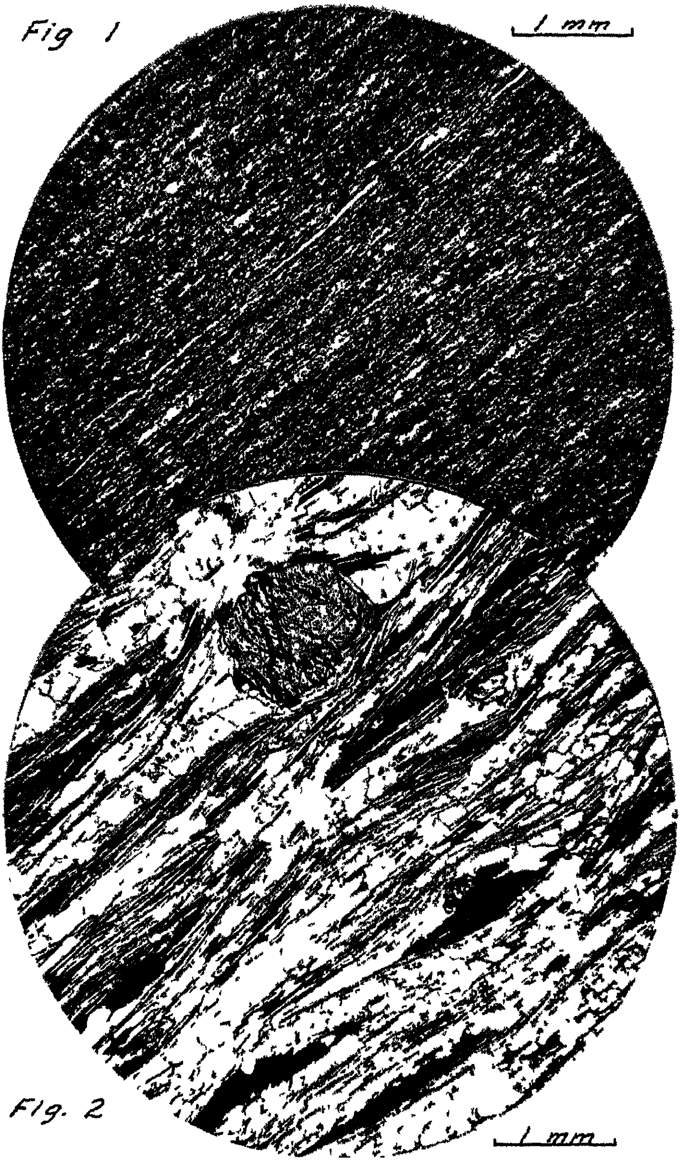


PLATE 1

Fig 1 Epizonally metamorphic argillite, phyllite Sericite and quartz Plain polarized light North of Kalapani, southeastern part of Nanga Parbat area (specimen no 16,821)

Fig 2 Mesozonally metamorphic argillite micaschist Muscovite, biotite, quartz, almandite Plain polarized light Rampur ridge, eastern part of Nanga Parbat area (no 19,625)

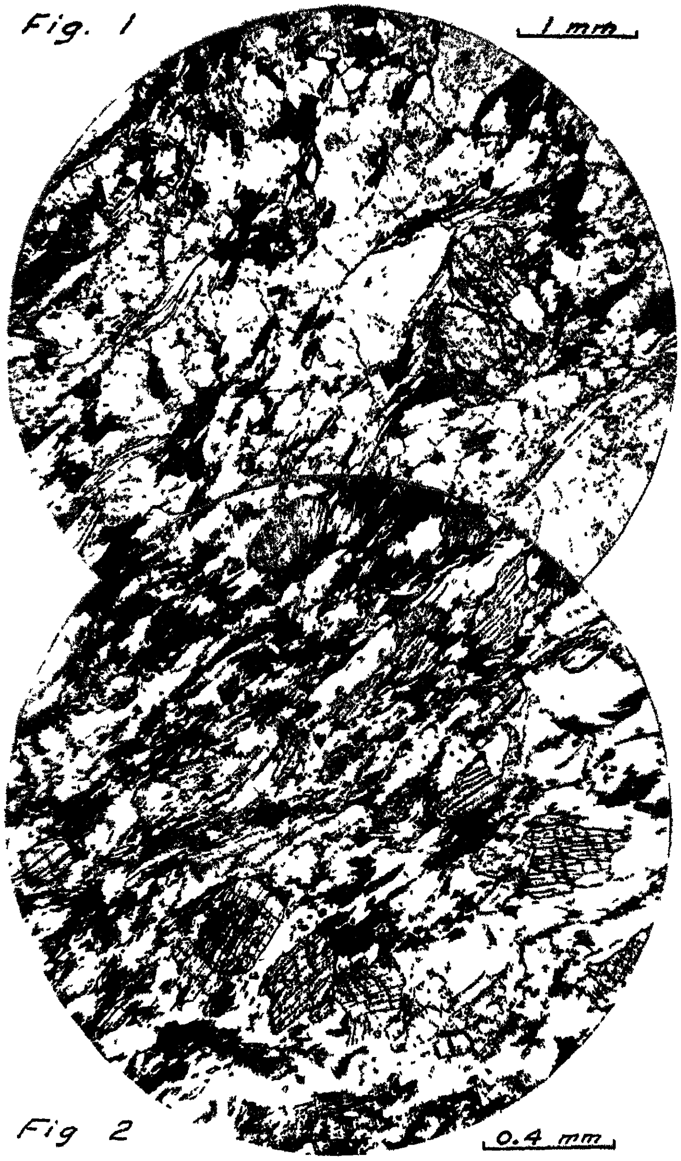


PLATE 2

Fig 1 Mesozonally metamorphic argillite biotite-paragneiss Biotite, muscovite (less), oligoclase, quartz, almandite Plain polarized light. Bunar Nullah, northwestern part of Nanga Parbat area (no 245).

Fig 2 "Low"-mesozonally metamorphosed argillite kyanite-biotite-paragneiss Kyanite, biotite, oligoclase, potash-feldspar, quartz, almandite, traces of muscovite Plain polarized light Ridge south of Rama Valley, eastern part of Nanga Parbat area (no 1262)

Fig. 1

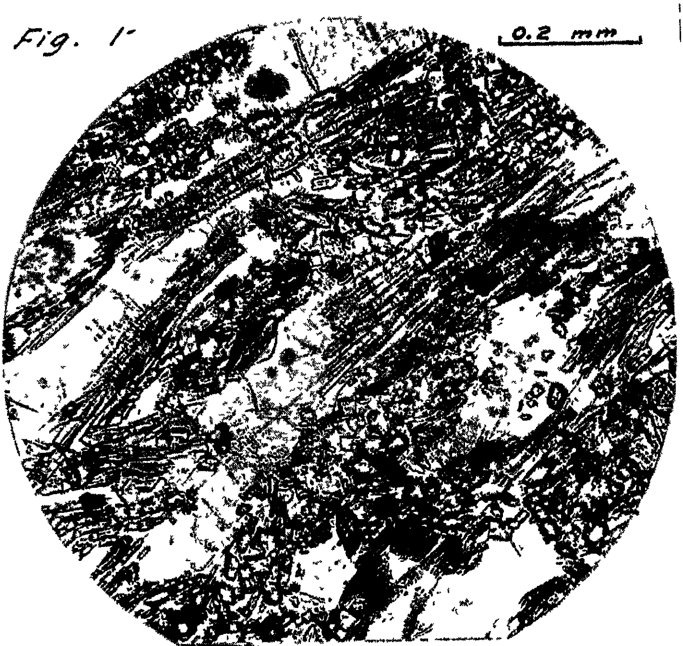


Fig. 2

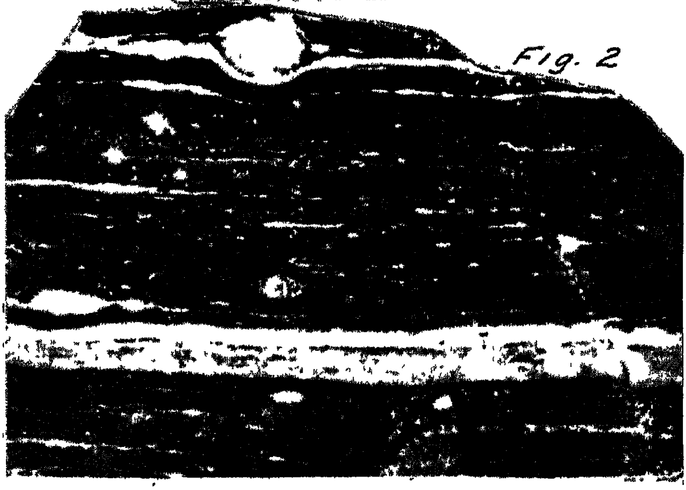


PLATE 3

Fig 1 Katazonally metamorphic argillite: sillimanite-biotite-paragneiss. Quartz-less variety, portion rich in sillimanite Sillimanite, biotite, orthoclase (sanidine), oligoclase (less). Plain polarized light North face of Nanga Parbat above Ganalo Glacier, central part of Nanga Parbat gneiss massif (no 24 7 8)

Fig 2 Beginning of granitization in metamorphic argillite mesozonal banded feldspar-porphyroblast-gneiss Rock surface, nat. size South of Mangdoyan north of Astor, eastern part of Nanga Parbat gneiss massif (no 13 5 8a).

Fig 2

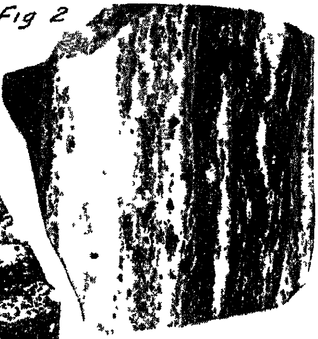


Fig. 1

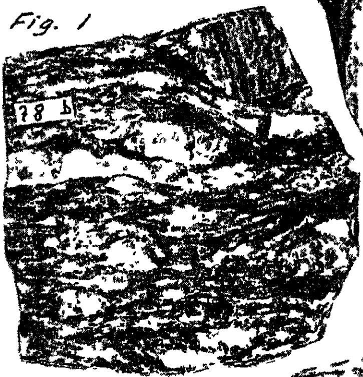


Fig 3

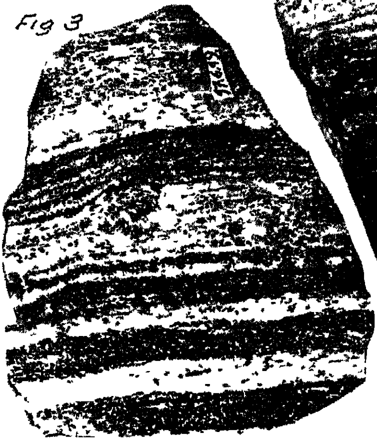


Fig 4

PLATE 1

Granitization in mesozonal metamorphic argillites

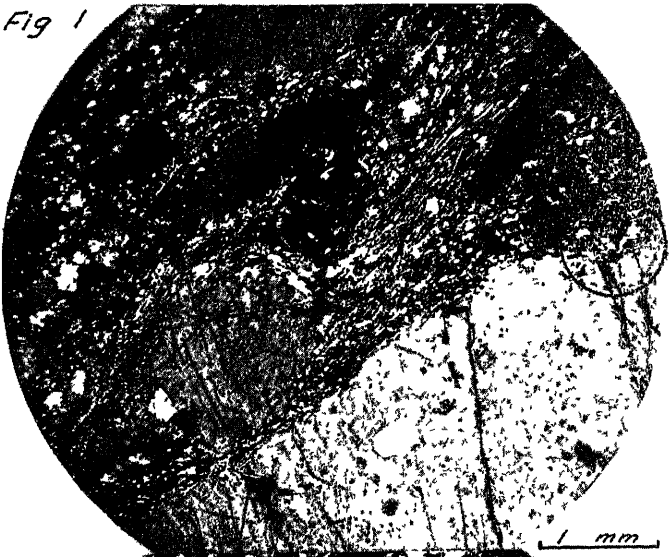
Fig 1 Augengneiss Potash-feldspar porphyroblasts replacing fine-grained dark colored kyanite-biotite-paragneiss Abt ½ nat size South of Astor, eastern part of Nanga Parbat gneiss massif (no 7 8b)

Fig 2 Banded gneiss (tourmaline-bearing) Completely granitized white quartz-feldspar bands, non-replaced dark-colored biotite-paragneiss bands, and intermediate bands Thinning of white replacement bands Small potash-feldspar porphyroblasts About ½ nat size North of Astor, eastern part of Nanga Parbat gneiss massif (no 9 6 5)

Fig 3 Banded gneiss Gradations between light-colored replacement layers and dark-colored biotite-paragneiss bands In replacement layers some porphyroblastic texture megascopically visible About ½ nat size Chugam area south of Astor, eastern part of Nanga Parbat massif (no 17 6 19)

Fig 4 Banded gneiss showing passage to augengneiss Potash-feldspar porphyroblasts in part coalesce into replacement bands About ½ nat size Gishat Nullah, southeastern part of Nanga Parbat area (no 17 8 8a)

Fig 1



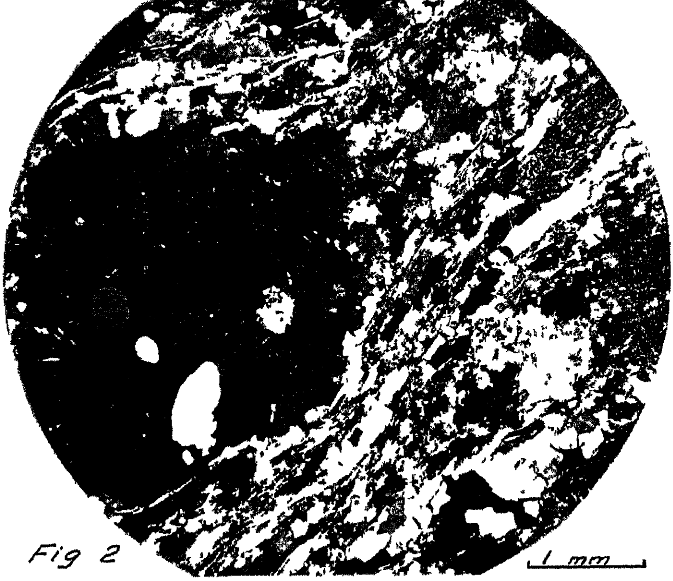
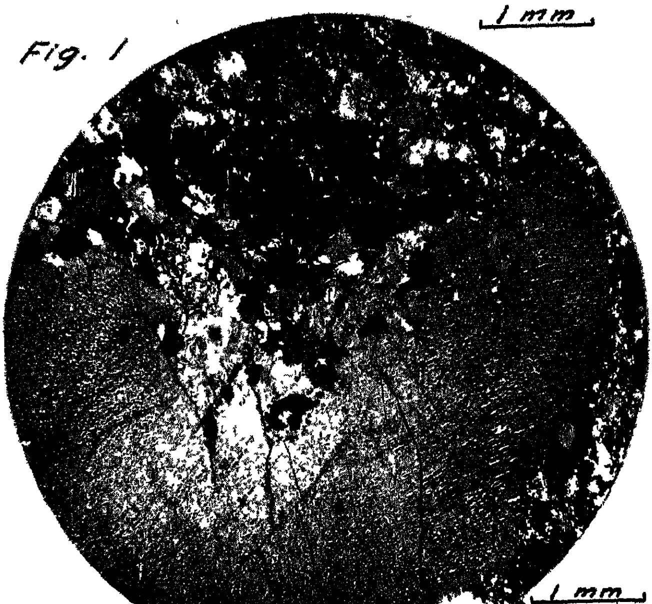


Fig 2

PLATE 5

Fig 1 Mesozonal augengneiss Microperthite porphyroblasts metasomatically replacing kyanite-biotite-paragneiss Crossed nicols Circle on right area shown in plate 7, fig 1 South of Astor, eastern part of Nanga Parbat gneiss massif (no 10 6 augengn -valley)

Fig 2 Potash-feldspar porphyroblast replacing kyanite- and biotite-bearing groundmass in incompletely replaced light-colored banded gneiss of mesozone Crossed nicols Ridge north of Rama Valley north of Astor, eastern part of Nanga Parbat gneiss massif (no 11 6 9)



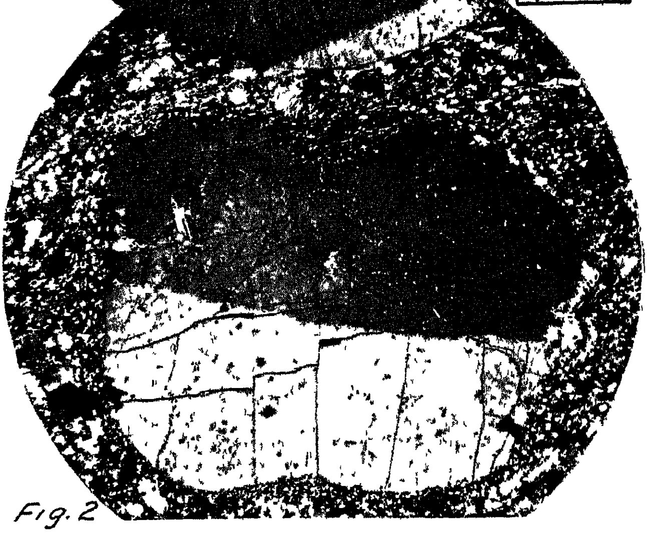
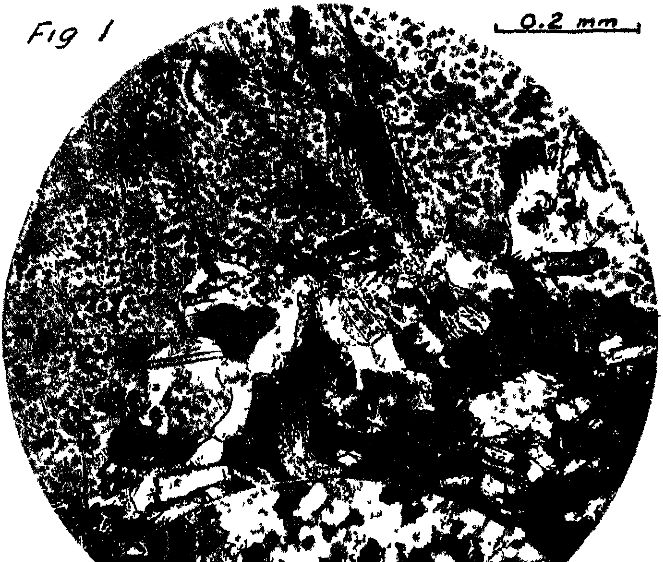


PLATE 6

Fig 1 Part of large microperthite porphyroblast replacing fine-grained kyanite-biotite-paragneiss in porphyroblastic and partly lenticular banded gneiss of mesozone Crossed nicols South of Astor, eastern part of Nanga Parbat gneiss massif (no 7 8 g)

Fig 2 Microperthite porphyroblast replacing very fine-grained biotite-paragneiss Mesozonal augengneiss comparatively poor in augen (passage to feldspar-porphyroblast-schist) Idioblastic tendency of porphyroblast Crossed nicols South of Astor, eastern part of Nanga Parbat gneiss massif (no 7 8 d)

Fig 1



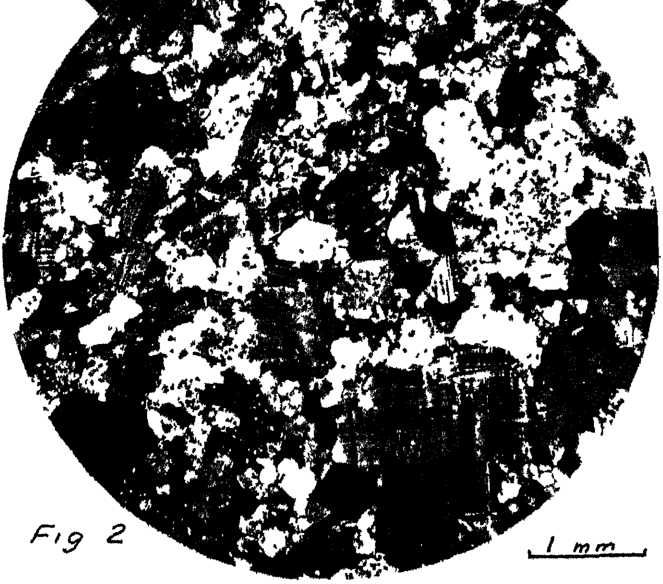


Fig 2

PLATE 7

Fig 1 Border of microperthite porphyroblast replacing kyanite- and biotite-rich paragneiss matrix in mesozonal augengneiss Detailed view of area marked by circle in pl 5, fig 1 Crossed nicols South of Astor (no. 106 augengn -valley)

Fig 2 Completely granitized white band in mesozonal banded gneiss Porphyroblastic microcline, quartz, oligoclase, very little muscovite Granoblastic Crossed nicols. Diehl Finger, eastern border zone of Nanga Parbat gneiss massif north of Astor (no 3737)

Fig. 1

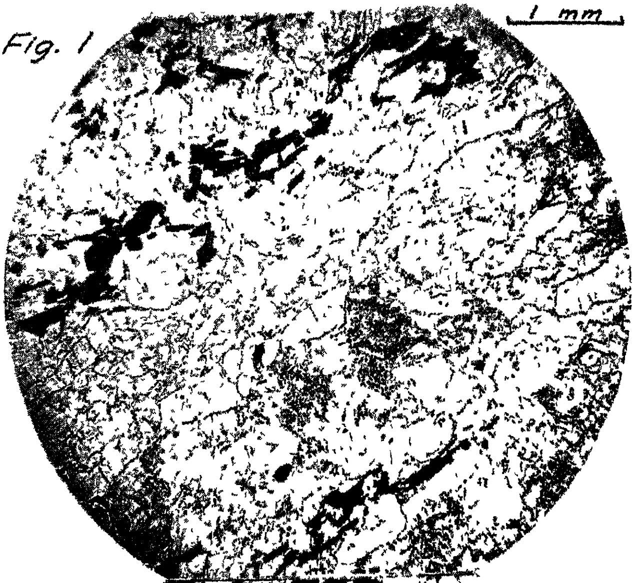


Fig 2

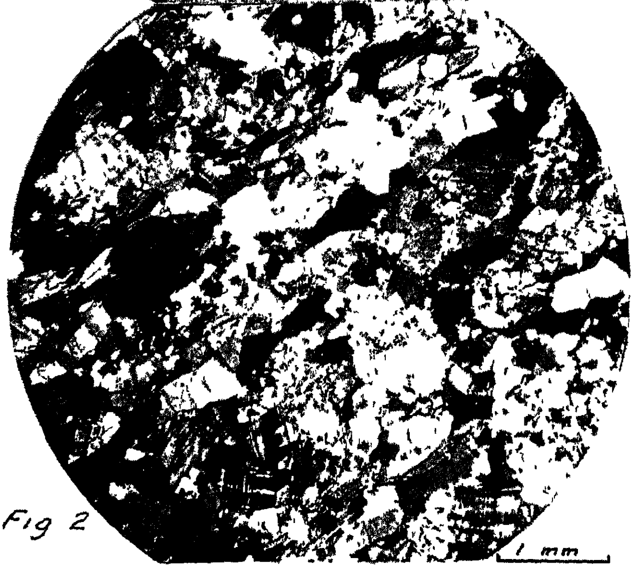


PLATE 8

Figs 1 and 2 Mesozonal banded gneiss Portion showing high degree of replacement Relict bands of biotite-paragneiss being invaded and absorbed Olive-green biotite, microcline, quartz, oligoclase, in biotitic bands accessory epidote with orthite Figs 1 and 2 show the same area, in plain polarized light and under crossed nicols East ridge of P 16,000′ west of Rattu valley, eastern part of Nanga Parbat gneiss massif (no 21 68)

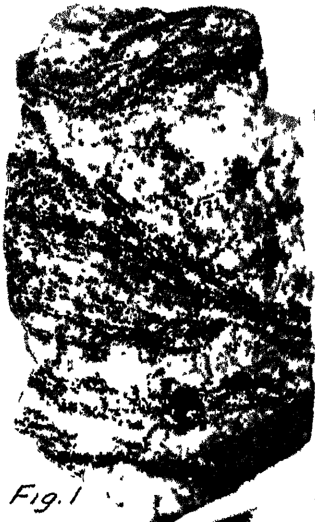


Fig. 1

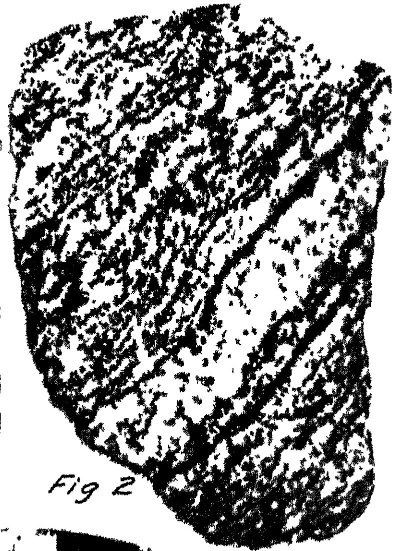


Fig 2

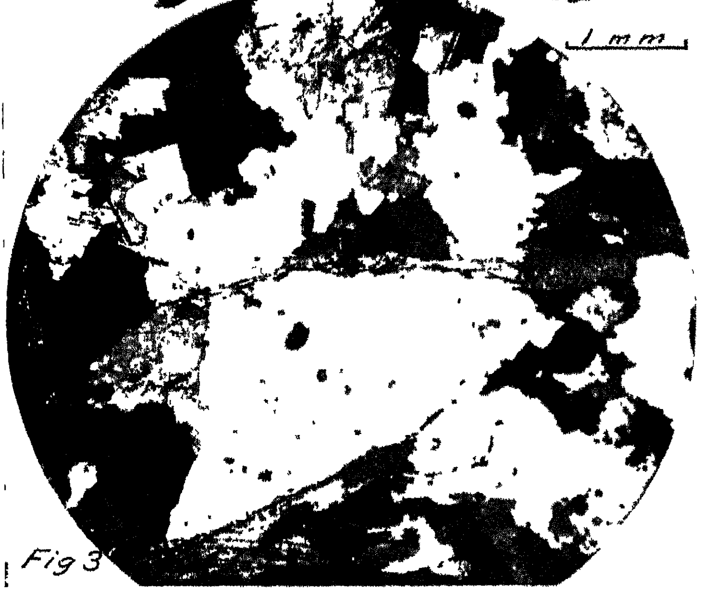


Fig 3

PLATE 9

Katazonal granitic gneiss forming interior of Nanga Parbat gneiss massif

Fig 1. Very coarse-grained biotite flasergneiss. Uneven texture. Porphyroblastic growth of feldspar About ½ nat size. Buldar ridge north east of Nanga Parbat (no 8 7 2).

Fig 2 Banded biotite-flasergneiss About ¾ nat size. North face of Nanga Parbat (no. 18.7 2e).

Fig. 3 Coarse-grained biotite flasergneiss Orthoclase, oligoclase, quartz biotite, some almandite. Final stage of granitization. Granoblastic texture Crossed nicols North face of Nanga Parbat (no. 18 7 1).

(pl. 4, fig. 3), and more generally visible under the microscope (pl. 5, fig. 2; pl. 6, fig. 1; pl. 7, fig. 2). At the borders of light-colored bands, the feldspars can often be seen invading the schist (in a manner comparable to that observed in the augengneisses), and thereby to "dilute" the schist so that its biotite becomes more thinly spread (pl. 8). Occasionally, more coarsely recrystallized biotite becomes concentrated in extremely thin layers (almost films) along sharp borders of white bands.

The similarity in composition of the dark-colored bands and the metamorphic Salkhala argillites of non-granitized areas, and the great variability in the thickness and proportion of dark- and light-colored bands compared with the slightly variable composition of the non-granitized schists, prove that *the banding cannot be due to metamorphic differentiation* of the argillite-derived country rock, but has been brought about by an introduction of material. The transformation of the metamorphic argillites into banded gneisses has been accompanied by a *definite change in the chemical composition of the rock*, and at the end of this series of transformations occur those banded gneisses in which unaltered schist material has become very subordinate. Metamorphic differentiation has taken place only very locally and on a very minor scale, producing those thin biotite-enriched bands which have been occasionally segregated at the borders of white bands.[14] Judging from the mineral composition of the rocks, the introduced material in the banded gneisses is alkali (predominantly potash), and usually also silica. The possibility that some sodium had been introduced already at an earlier stage of metamorphism, has been discussed above (cf. chapter 1).

Which was the *mechanism of introduction* of this material? Banded gneisses of the type described are commonly considered as the result of "*lit-par-lit injection.*" There are considerable divergencies in the use of the term "injection." In this country, it is usually applied in its restricted and proper sense, e. g. for an intrusion of magma or magmatic solutions or similar mobile materials into some space created by a breaking-open or splitting-up of the solid country rock, emplacement being a mechanical process. In Europe, the term "injection" has been widely used in a looser and more comprehensive sense, comprising processes of infiltration, soaking and replacement, as well as true injection, but excluding intrusion of larger continuous bodies of magma. The term has, for instance been used in this sense by V. M. Goldschmidt in his classical study of the "injection metamorphism in the Stavanger region" in Norway (Goldschmidt, 1921), in which part of the "injection gneisses" described are shown to owe their origin to infiltration and replacement by solutions (assumed to be derived from a magmatic source). More recently, C. E. Wegmann has, in an important contribution to the analysis of migmatites (Wegmann, 1935), suggested the restriction of the term "injection" to its proper meaning. The definition given by him agrees with the usage customary in this country; with the modification that intrusion of larger continuous bodies of magma is excluded from the term "injection." In a preliminary report on the Nanga Parbat area written immediately after my return from the field (Misch, 1935), I have used the term "injection" in the broad sense which used to be customary in Europe, but I would now apply it only in its restricted sense. In that paper, I applied the term "injection gneisses" to various types of gneisses of metamorphic-migmatic origin, including the porphyroblast-gneisses and the massive augengneisses, although their replacement origin was pointed out in that paper. The banded gneisses I then considered as the result of true lit-par-lit injection by acidic-granitic material derived from a magmatic source. Subsequent petrographic study of the rock suites from Nanga Parbat has, however, convinced me that most, and probably all, of the material composing the light-colored layers in the banded gneisses owes its origin to *infiltration and replacement of the country rock, rather than to true injection.*

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[14] In the Nanga Parbat area, metamorphic differentiation appears to be responsible for the formation of certain banded amphibolites derived from some of the norites and related rocks. In this case, lighter and darker bands are composed of exactly the same mineral varieties, and only their relative amounts differ in different bands. There is no indication that the bulk composition of the rock has been changed. The individual bands are uniformly thin, and the proportion of lighter and darker bands does not vary greatly I consider these features as characteristic of banding originated by metamorphic differentiation, in contrast to banding due to introduction of foreign material.

The following features which have been described above, are in favor of such an interpretation by "*lit-par-lit replacement*": the presence of numerous bands intermediate in composition between the original schists and the white gneiss bands, and the occurrence of gradational contacts—features which suggest different degrees of replacement of the original schist; the intimate association of banded gneiss with, and its grading into, undoubted replacement rocks, e.g. augengneiss and feldspar-porphyroblast-schist; and the porphyroblastic texture widely developed in light-colored gneiss bands, which links such bands still more closely to layers of porphyroblasts that often prolong the light-colored bands. There is also a general argument against an origin of the banded gneisses by true lit-par-lit injection. It is difficult to visualize that schists while being subjected to considerable pressure and to action of intense orogenic stress, could have been split open by magma or magmatic solutions throughout areas measuring tens of miles along the strike and a number of miles across it. On the other hand, infiltration along layers of active shearing could have taken place under such mechanical conditions. This argument seems to be even more convincing where banded gneisses do not dip steeply, but have gentle dips, as in certain parts of the Nanga Parbat area,[15] and as in some other regions I have since had an opportunity to study. If in such cases banding were the result of true injection, each injected layer of liquid would have had to uplift and carry the tremendous load of thick overlying rock masses.

The suggestion that lit-par-lit replacement has played the main part in the formation of the banded gneisses, does not preclude the possibility that there may have been some subordinate contribution of true injection material which became activated as a final result of the migmatic metamorphism. However, no structural features indicative of mass movement of granitic material were observed. On the contrary, the tectonic structure of the banded gneisses—as well as of all other migmatitic gneisses in the Nanga Parbat area—is of a type characteristic of deformation of solid rocks during orogenic metamorphism, and is identical with the type of structure seen in the non-migmatitic schist areas. This general structural aspect of the problem will be discussed below.

Lit-par-lit replacement, as against true lit-par-lit injection, implies an emphasis on *chemical processes*, as against a purely mechanical process. The introduced material infiltrated and permeated the country rock, and reacted with it. To make banded gneiss, the *loci of infiltration and reaction* must have been thin layers in some way distinguished from neighboring layers. One might think of primary differences in the mineral composition of the schists, certain layers being more pervious or having a greater chemical affinity to the granitizing solutions.[16] However, the areas of non-migmatitic metamorphosed argillites do not show primary differences of this kind, and in the southeastern part of the Nanga Parbat region these areas of metamorphosed argillites grade into banded gneisses and augengneisses along the strike (cf. fig. 1). The explanation must then lie in some other condition that existed during the time of migmatitic metamorphism. It appears that this was the condition of active shearing movements along the foliation planes of the country rock which took place during the regional metamorphism and the migmatitization connected with it. The introduced material, consisting of solutions,[17] infiltrated the country rock *along layers distinguished as particularly active shear zones* during the differential movements which were taking place throughout the country rock. The loci of layers of feldspar porphyroblasts were determined in the same manner as those of the white bands eventually formed.

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[15] Parts of the southeast and northeast; cf figs. 17 and 21 in (Misch, 1935).

There arises the question why the result of migmatitic infiltration should have been predominantly massive augengneiss in certain marginal belts, and chiefly banded gneiss in neighboring zones. These two rock varieties cannot be considered merely as less and more advanced stages in the process of migmatitic transformation; for the banded gneisses contain layers of non-migmatitic schist whereas such are, apart from subordinate intercalations, absent in the augengneiss belts. Incipient banded gneisses are represented by schists with layers of porphyroblasts, but not by uniformly feldspathized rocks. The difference between the belts of augengneiss and the zones of banded gneiss are probably due to a more highly volatile character of the infiltrating solutions in the marginal belts predominantly transformed into augengneiss, the solutions thus being able to permeate the country rock in a more uniform manner. Such a disposition seems plausible at a time when the "migmatite front" (to use a term coined by Wegmann, 1935) had advanced to its maximum extent and remained more or less stationary for a period.

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[16] The latter argument applies wherein the Nanga Parbat massif calcareous or basic intercalations have, in contrast to the enclosing metamorphic argillites, escaped granitization.

[17] As to the solutional character of the introduced material, I refer to the discussion with respect to augengneiss.

It has been stated above that the banded gneisses were formed when orogenic stress was active, which means under conditions of differential movement within the rock masses concerned. Foliated structure within the light-colored as well as the schist bands, general sharp parallelism of banding and foliation and of other elements such as stringers of porphyroblasts, and presence of some cataclastic mineral structures not healed, prove the *synkinematic* (synorogenic) character of the banded gneisses.

(*b*) *Migmatitic gneissose granites in the interior of the Nanga Parbat massif*

Though the augengneisses and banded gneisses in the outer zones of the Nanga Parbat gneiss massif have been shown to owe their origin to metasomatic granitization, it is still feasible that these zones, though having a width of miles and occupying an area of more than 100 square miles, could be nothing but extensive contact zones granitized along the margins of a synorogenic intrusive body occupying many hundreds of square miles in the interior of the massif. We cannot therefore speak of "metasomatic granitization of batholithic dimensions," unless it also can be demonstrated that the granitic gneiss in the interior of the massif owes its origin to granitization, and not to igneous intrusion.

The rocks forming the interior of the gneiss massif have been outlined in chapter (II), and the non-granitized metamorphic sediments occurring here briefly described in chapter (III). Whereas in the banded gneisses and augengneisses of the marginal areas large quantities of argillite-derived fine-grained paragneiss and biotite-schist material are preserved and are easily distinguished from the metasomatically replaced portions of the rocks, this distinction becomes less well marked in the interior of the massif, and typical fine-grained paragneiss material is preserved only in quantitatively very subordinate intercalations. Otherwise, the argillite-derived rocks gradually become more coarsely recrystallized and feldspar-enriched, as the interior of the massif is approached. The result is mainly flasergneisses of moderately fine to medium grain and of medium shades. Their sedimentary derivation is only shown by their grading into typical fine-grained paragneisses (cf. Misch, 1935). Toward the interior of the massif, these rocks pass into the *coarse-grained biotite-flasergneiss of granitic appearance and composition* (*or gneissose granite*) which here predominates (cf. pl. 9, figs. 1 and 2).

This gneiss is composed of orthoclase, oligoclase (varying from a little over 10 to somewhat over 20 percent anorthite), quartz, biotite, and frequently also muscovite as an equal and contemporaneous constituent. Almandite is a common accessory. Although the rock is rather uniform regionally, it is anything but homogeneous in detail, and its *texture is usually very uneven.* The gneiss is generally coarse-grained though the grain size ranges from more medium-grained to extremely coarse-grained varieties, often within very short distances. Some of the gneisses are even-granular, but an uneven grain, due to the development of large feldspars (especially orthoclase), is more common and occurs always in extremely coarse-grained varieties. The biotite content of the gneiss varies greatly, even within short distances. In single rock specimens the biotite is usually unevenly distributed, being concentrated in bands, stringers and patches (pl. 9, figs. 1 and 2). The almandite may also be enriched in certain bands. The gneisses possess crystallization-foliation (free of cataclastic mineral deformation), but the degree of their schistosity is variable and never becomes extreme. Most of the rocks show flaser structure which may pass into augen structure (cf. pl. 9, fig. 1). More typical augengneisses are subordinate. They differ in their texture from those of the marginal parts of the Nanga Parbat massif, their groundmass being more coarse-grained and their orthoclase porphyroblasts usually being less well-formed crystals. In some augengneisses, the biotite-rich groundmass exceeds the porphyroblasts. Banded structures are quite common in the gneisses, and are often found in light-colored varieties (pl. 9, fig. 2). Acidic bands are occasionally bordered by thin layers enriched in biotite (local metamorphic differentiation). The banded and augen structures described recall those typical of the marginal zones of the massif, with coarse-grained crystallization of the entire rock and more general feldspathization superimposed. Banded and augengneisses of the type of the marginal zones, characterized by the preservation of fine-grained and non-granitized schist material, are very rare in the interior of the massif; they occur only occasionally in association with the subordinate intercalations of fine-grained paragneiss found in the granitic gneiss. The microscopic texture of the granitic gneiss is crystalloblastic and usually shows the same uneven character which is megascopically visible; it can be described as granoblastic, partly with porphyroblastic tendencies. Many gneisses show a tendency to approach an almost igneous-appearing microscopic texture which many petrographers would classify as "pan-allotriomorphic granular," but the crystalloblastic character and the uneven nature of the fabric remain preserved. An advanced case is shown in plate 9, figure 3.

The character of the gneiss is strongly suggestive of a metamorphic-migmatitic rather than an intrusive origin. Among the features described, the following are especially indicative of such an origin: variability of composition, grain size and other textural features; uneven nature of texture, frequent presence of porphyroblastic features; crystalloblastic texture (though this latter does not constitute conclusive evidence since it could be superimposed). Of particular interest are the augen and banded structures described above; they suggest that these rocks had originally gone through the stage of mesozonal incomplete granitization which is now displayed in the marginal zones of the massif, and had subsequently, with further spreading of granitization, become more coarsely recrystallized and more thoroughly feldspathized, so that the paragneiss material still preserved in the mesozonal banded and augengneisses lost its distinctive character, although the banded and augen structure is preserved as a relic. In general, the biotite-rich stringers and patches in the unevenly textured granitic gneisses are rather suggestive of recrystallized relics of biotite-paragneiss. Exceptionally, sillimanite was found to occur in granitic gneiss, indicating participation of argillite-derived material.

Apart from the character of the gneiss itself, there are *other and even stronger arguments in favor of an origin by metasomatic granitization.* The first is the presence of *intercalations of undoubted metamorphic sediments* and some basic rocks which, though quantitatively subordinate, occur throughout the wide area of granitic gneiss. *These intercalations are structurally conformable with the granitic gneiss.* Most of them are thin, and many of them can be followed for long distances. It is difficult to visualize that mechanical intrusion of magma could have spared such septa and, even if it did, could have left them structurally intact without thoroughly disturbing and disrupting them. If, on the other hand, these intercalations are interpreted as relict bands which have escaped metasomatic granitization, their presence and their mode of occurrence are readily understood.

Many of these intercalations consist of katazonal fine-grained paragneiss whose contacts with the granitic gneiss are usually gradational. Some layers within such paragneiss intercalations may be partially granitized, having been transformed into feldspar-porphyroblast-schist and porphyroblastic augen-gneiss. Often paragneiss and granitic gneiss alternate in parallel layers. All these features indicate granitic replacement rather than intrusion.

Of particular interest are the *thin marble layers* with their associated lime-silicate-granulite bands, which form regular conformable intercalations in the gneiss often traceable for long distances. None of the marble layers observed exceed a thickness of 100 feet and most of them are much thinner. No intrusion could have left these bands structurally intact; moreover, it is difficult to comprehend how the marbles could have escaped fusion if temperature had been high enough to keep silicate compounds in a molten state at the existing pressure for a considerable time in large magmatic masses surrounding the marbles. There is more marble relative to metamorphic argillite in the interior of the gneiss massif than in the area of non-granitized Salkhala phyllite and of schist and paragneiss derived from it. The absolute proportion of marble, on the other hand, is nearly of the same order in the interior of the massif as in the non-granitized areas. This proves that most of the argillites which greatly predominate in the original sedimentary section, became transformed into granitic gneiss whereas most, if not all, of the marbles escaped granitization—a truly metasomatic process, and one of differential character. In this connection, it should be recalled that the non-granitized Salkhala argillites grade into the mesozonal migmatitic gneisses and then into the central granitic gneisses not only across but also along the strike. Generalizing, it may be said that the marbles extend from an area of phyllites through one of schists into one of mesozonally migmatitized schists and continue right into the granitic gneisses. The relict nature of the thin marble bands is particularly striking where they occur as conformable intercalations in coarse-grained granitic gneiss without being accompanied by any paragneiss, all of the argillite originally enclosing the marble having been transformed into granitic gneiss. Several cases of this kind have been observed (cf. fig. 2), although the majority of the marbles are still associated with some paragneiss which varies in thickness from a few inches to several hundred feet.

The reason why calcareous bands have escaped granitization is obvious. Whereas it does not take much introduced material to transform an argillite into a rock of granitic composition, the substance of a marble would have to be almost entirely replaced to achieve the same result. Apart from the different chemical affinity of marble and argillite to granite, their relative permeability to alkali-rich solutions is probably an additional factor. Whereas slate is undoubtedly less pervious to aqueous solutions than limestone, their metamorphic equivalents must be supposed to act differently under conditions of granitization. Now metamorphic argillite would be more pervious, consisting of a granular fabric of silicate minerals (and being, in case of synkinematic granitization, in a state of active differential movement); and, having an excess of aluminum, metamorphic argillite would have a strong tendency to precipitate alkali from permeating solutions rich in alkali. On the other hand, under the existing high pressure preventing escape of its CO_2, marble would be less pervious to those solutions; at the same time, it would have no tendency of precipitating alkali from solutions rich in it. Also, under pressure, marble might plastically flow and hence be less permeable than schist traversed by active shearing planes.

Basic rocks—chiefly katazonal amphibolites—intercalated in the interior of the gneiss massif, and originally forming bands in the Salkhala argillites, have likewise essentially escaped granitization. Generally, granitization of basic rocks is not uncommon, and has been described by a number of authors. In areas where I have observed this process, there is usually the following series of transformation · original basic rock (amphibolite, basic hornfels, etc.; sometimes with lime-silicate rock at the beginning of the series, other times with metabasalt, greenstone, metanorite, metagabbro, etc.); gabbroid rock; dioritic rock; quartz-dioritic and, not always, granodioritic rock (for instance: Indus Valley north of Nanga Parbat; Fish Lake, Palmer Lake and other places in Okanogan country, Eastern Washington; Northern Sierra Nevada; all these examples are postkinematic). In this series, the alkali added apart from silica etc. is sodium—in contrast to the predominant potash metasomatism in the Nanga Parbat gneiss massif—and as far as potash has at all been introduced, it has come in only near the end of the transformation. There can be no doubt that there are two chief chemical types of granitization, i.e. *predominant potash metasomatism,* and *predominant sodium metasomatism.* Both are apt to act on metamorphosed argillites, but the latter seems to be more easily able to attack basic rocks apart from argillites. The only effect of granitization of the potash metasomatism type on basic intercalations which I have observed in the Nanga Parbat gneiss massif, consists of biotitization of hornblende which has often marginally transformed hornblende-rich intercalations into almost pure schistose biotite rock.

Summing up from a structural angle, the relationships described can be said to be characterized by *structural continuity.* This applies if we proceed from the non-granitized zones through the incompletely granitized to the more fully granitized areas, and is also valid in a stratigraphic sense, insofar as individual stratigraphic units run through these three zones. Structural continuity also exists within the granitized areas between granitized rocks and non-granitized relics (regularity of mutual intercalation, conformable and gradational relations, harmonic and equal participation in the same individual folds).

The structural features listed confirm the replacement origin of the granitic gneiss, and illustrate the absence of intrusive structures. The same is demonstrated by the *type of tectonic deformation.* It is identical in the zone of phyllites, the area of non-granitized schists, the mesozonal migmatitic belts, and the katazonal region of granitic gneiss with its relict intercalations. The structure in all these zones is characterized by well-defined trends which remain constant for long distances and are clearly related to the regional structure of the northwest Himalayas, and by a regular character of the major folds. There is no sign of that structural turmoil and turbulent folding and flowage which one would expect if magma had broken into the sediments under high pressure and intense stress, and which is known to occur even in solid rocks deformed under conditions of extreme plasticity. There are, of course, as in all areas of crystalline schists, numerous intensely compressed minor folds of "mobile" character throughout the zones described, but they are, in the interior of the gneiss massif as well as in the phyllite area, all regularly aligned parallel to the prevailing local trend. Most of these minor folds are isoclinal, and their narrow arcs can often be seen to become obliterated by the foliation which is parallel to their limbs.

The type of structure described is that *typical of solid rocks deformed under conditions of regional metamorphism.* The fact that this type of deformation characterizes the granitic gneiss area as well as the phyllite zone, is the last argument I wish to offer in favor of a replacement origin of the granitic gneiss or gneissose granite, and therewith of the whole of the vast gneiss massif of Nanga Parbat.

A short discussion on the *chemical composition of the "granitizing solutions"* is necessary. The *predominant solvent* was undoubtedly water (presence of hydrous silicates in all metamorphic zones; water content of the original rocks). As to the substances carried, *material present in a solution* and *material precipitated from a solution* must be distinguished. Only the latter is *directly recorded* in the rocks. If, for instance, an aluminum-excess rock has precipitated potash from a solution and augengneiss has thus been formed, the permeating solution may have contained other substances not recorded by precipitation. However, certain *indirect conclusions* are possible. The solution from which potash was precipitated must have had an excess of alkali over aluminum, if it did at all contain more than traces of aluminum. Moreover, the solution must have contained more potash than sodium. Otherwise, the porphyroblasts of microperthite in the augengneiss would be accompanied by such of sodic plagioclase. On the other hand, the solution may have contained a large or a smaller amount of silica. Silica would not be necessarily precipitated by a rock already possessing an excess of free silica. When kyanite and mica of the original schist were replaced by feldspar porphyroblasts, some silica was consumed. Some silica was supplied by quartz of the original schist, which together with mica and kyanite was replaced by feldspar. Where the amount of quartz was not sufficient, additional silica was precipitated from the solution. Whether a small amount of silica has been added in the augengneisses, could be ascertained only by a series of chemical analyses. In the white replacement layers in the banded gneisses silica has undoubtedly been introduced in addition to alkali. For these layers are rich both in feldspar and quartz, and poor in mica, and thus more silicic than the original mica-rich schist.

Another indirect conclusion as to the composition of the granitizing solutions can be drawn from the plagioclase in the granitized rocks. The mesozonal augengneisses and banded gneisses usually contain sodic oligoclase, like the non-granitized metamorphic argillites. If the granitizing solutions had contained any considerable amounts of calcium, the plagioclase in the mesozonal granitized rocks would have become a calcic oligoclase, and possibly epidote minerals would have formed. Even the katazonal migmatitic granitic gneisses usually contain a sodic oligoclase. If the granitizing solutions had contained even a moderate amount of calcium, the plagioclase would of necessity have become more calcic under katazonal temperatures. If the solutions had had any considerable content in iron (and magnesium), the amount of biotite could not show a marked decrease in the granitized rocks compared to non-granitized schists, and the intercalated marbles would probably have been affected.

This leads to another criterion. Where granitizing solutions have acted *differentially* on rocks of different composition, further indirect conclusions as to the chemical nature of the solutions are possible. If in the same section rock A has been granitized and rock B has escaped granitization, the permeating solutions were obviously rich in those substances required to transform rock A into one of granitic composition, but deficient in those elements primarily needed to convert rock B into one of granitic composition. Solutions which were "granitizing" with regard to rock A, do not deserve this name with regard to rock B. Differential granitization therefore makes it possible to eliminate a number of substances as essential constituents of the solution.

An example is granitization of basic rocks, such as amphibolites, in comparison to that of metamorphic argillites. It has been shown above that basic rocks are mainly apt to be granitized if the permeating solutions have a high sodium content—apart from the large amount of silica which is of course required. It seems that predominant potash metasomatism does not generally affect basic rocks very much (apart from biotitization), although potash is frequently introduced at a late stage of the transformation of basic rocks, whereas the earlier stages are usually characterized by definite sodium predominance.

(IV) RELATIONS BETWEEN PROGRESSIVE REGIONAL METAMORPHISM AND SYNKINEMATIC GRANITIZATION

From a general point of view, one of the most striking features displayed in the Nanga Parbat area is the *intimate relationship and the simultaneous increase of progressive regional metamorphism and of synkinematic granitization.* Throughout the area, granitization does not begin before regional metamorphism in the argillites has reached the stage of the low-mesozonal kyanite-schist facies, and it invariably does start in this zone. It here remains rather incomplete—considerable quantities of non-replaced argillite material being preserved—and its type is that represented by banded and augengneisses. As regional metamorphism approaches the stage of the katazonal sillimanite-paragneiss and lime-silicate-granulite facies, granitization becomes more complete—argillite relics being subordinate—and develops the type represented by granitic gneiss or gneissose granite.

The succession of zones and mineral facies of regional metamorphism can in this area be shown to be exclusively *a function of increasing temperature* (cf. chapter II a). Since the degrees of regional metamorphism and of synkinematic granitization are definitely linked, this latter must likewise increase with temperature, and the same must apply to the introduction of material which is implied in metasomatic granitization. In other words, *rise of temperature and increasing infiltration by granitizing solutions go parallel.*

The rise of temperature during metamorphism and granitization in this, and many other areas, is *unrelated to geologic depth,* i. e. is not a normal geothermal rise.[18] There must, therefore, have been a *differential supply of heat* resulting in an upward curving of geisotherms—at very steep angles in this area in which the metamorphic isograds are approximately vertical. There are no known processes which could have created a sufficiently large amount of heat *in situ.* Heat must therefore have been *introduced* from below. Was this done by conduction or convection? The former seems unlikely in itself, considering the high angle of upward curvature of geisotherms; the systematic relation between rise of geisotherms and introduction of granitizing solutions confirms that *convection was the means of introduction of heat.* In other words, *the granitizing solutions carried both chemical substances and heat.* It is in these terms that we have to conceive the "rising of the migmatite front," to use again the phrase coined by Wegmann (1935). If no regional introduction of material were involved in granitization, it could be argued that progressive metamorphism, being itself controlled by rise of temperature, is the cause and granitization merely a consequence. However, material has undoubtedly been added "from somewhere below"; to demonstrate this fact, the chemical changes involved in the making of the migmatitic gneisses have been stressed, and it has been pointed out why these gneisses cannot have formed by processes of metamorphic differentiation of the country rock, i. e. by local transfer as against regional introduction of material.

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[18] Normal geothermal rise of temperature can, of course, also reach temperatures sufficiently high for high-grade regional metamorphism, but *only at great depths* of the crust. Here, migmatitization may take place without introduction of additional heat.

If the analysis given above is correct, *high-grade regional metamorphism is not only necessarily linked with synkinematic granitization, but depends on it for a supply of additional heat, with the exception of great depths of the crust.* In the Nanga Parbat area, regional metamorphism without granitization did not go beyond the phyllitic (and perhaps uppermost mesozonal) stage. All zones of higher-grade metamorphism are in their spatial distribution so clearly related to the central rise of the migmatite front, that a genetic control is obvious. This applies not only to the high-grade metamorphic sediments preserved within the granitized area, but also to the zone of higher-grade rocks outside the margin of the granitized area. The rocks involved are the typically mesozonal (higher mesozone and beginning of lower mesozone) mica-schists and paragneisses which form a belt between the low-grade rocks and the mesozonal augengneisses and banded gneisses. To this belt, additional heat must have been supplied from the neighboring granitized area, and the means may have been conduction, since the rocks in question are outside the granitized area and heat could not therefore have been conveyed to them by the granitizing potash-rich solutions directly—unless this belt was, outside the area of actual granitization, reached by aqueous solutions carrying some sodium (cf. above). To generalize, *the front of synorogenic granitization is in time and space preceded by a zone of regional metamorphism of a higher grade than that achieved by orogenic metamorphism not linked with granitization.* It is obvious that *the further increase* (from low-mesozonal to katazonal) *of the metamorphic grade in the sediments preserved within the granitized area of Nanga Parbat, illustrates the control by successively increased introduction of heat, conveyed by successively more active and hotter granitizing solutions.* In the granitized rocks themselves, this progression is shown quantitatively by increasing completeness of granitization, and qualitatively by the passage to a more truly granitic type of gneiss in the interior of the massif.

The *Nanga Parbat example* of progressive metamorphism and synkinematic granitization is *typical* and, it may be said, *normal*; its only distinguishing features are unusually clear, deep and continuous exposures. The rocks and processes described are repeated all over the world. Provided the principles suggested are correct at Nanga Parbat, they will also apply elsewhere, subject, of course, to certain modifications due to varying physical and chemical conditions (cf. below). However, one qualification must be made. Synorogenic granitization and metamorphism at Nanga Parbat took place in a *geosynclinal environment.* Therefore, areas which can be properly compared are other geosynclines. Metamorphism and granitization are characteristic of orogenic transformation in the interior portions of first-order geosynclines, and they here often rise to *comparatively high levels. At extreme depths* of the crust, processes will be modified. Such cases are believed to be displayed in some Archeozoic regions, and perhaps in the roots of some exceptionally deeply eroded later geosynclinal chains. Numerous areas of Archeozoic crystalline schists do not, however, differ from such of later ages, and can frequently be shown to represent old geosynclines exposed at levels comparable to those visible in later geosynclines.

The statement made above concerning the Nanga Parbat area, that *orogenic metamorphism without granitization* is apt to achieve only low-grade mineral facies ("pure dynamometamorphism") is borne out by numerous migmatite-free regions of phyllites and basal-epizonal and upper-mesozonal micaschists, and of other low-grade schists (greenschists, etc.). Among regions I have studied, a good example is the geosyncline of West Yunnan at the southwestern border of China (cf. Misch, 1945 a and b). Here Paleozoic and Mesozoic sediments and volcanics have been dynamically metamorphosed during Mesozoic orogeny into epizonal and occasionally uppermost-mesozonal crystalline schists within an area of many thousands of square miles, without having achieved a higher grade of metamorphism anywhere. No synkinematic migmatites of Mesozoic age have been observed anywhere in this region.

The *beginning of granitization* of the potash-metasomatism type was at Nanga Parbat found to be within the lower mesozone, and granitization was shown to be preceded in space and in time by mesozonal regional metamorphism resulting from an introduction of additional heat from the area of granitization (possibly connected with some finely divided sodium-precipitation from solutions). These relations seem to be general; for in all areas of synorogenic migmatites known to me, metamorphism has achieved a mesozonal mineral facies before granitization has set in. However, there appears to be some variation as to the part of the mesozone, and therewith as to the temperature, at which granitization begins. There seem to be cases where migmatic processes start already higher within the mesozone. A good example seems to be that of the Stavanger region in Norway described by V. M. Goldschmidt (1921). Here synorogenic granitization is predominantly of the sodium-metasomatism type. It is entirely of this type in the zone of beginning replacement (albite- and oligoclase-porphyroblast-schists); this zone is upper-mesozonal, judging from the minerals described in the schists. It appears, from the rock succession recorded by Goldschmidt, that here additional potash-metasomatism came in only at a more advanced stage of progressive granitization.[19] There seem to be other

[19] The more typically granitic rocks in this region are described as intrusive, but the continuous nature of the series is so pronounced, and the general character of the rock succession so strikingly resembles that described in the present paper, that an identical genetic interpretation suggests itself.

cases which also suggest that synorogenic granitization of the sodium-metasomatism type is apt to begin at a temperature lower than that at which potash-metasomatism was found to commence at Nanga Parbat.

Finally, I wish to emphasize *the general validity of the statement that high-grade regional metamorphism and synorogenic granitization are invariably and necessarily linked.* They are expressions of one and the same process. There is not any region known to me either from study of the literature or from personal field work, where katazonal or lower-mesozonal crystalline schists occur without metamorphic migmatites or granitic rocks.

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(*To be Continued*)

REGULARITY OF OLD FAITHFUL GEYSER, YELLOWSTONE NATIONAL PARK, WYOMING[1]

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ABSTRACT. It is common belief that old Faithful Geyser, the most celebrated geyser in the world, (1) has become irregular and even erratic, (2) does not erupt in winter, (3) erupts to heights much less than formerly, (4) discharges much less water than formerly, and (5) is the only regular large geyser.

This erroneous belief is based upon (1) inaccurate observation, (2) incorrect interpretation and correlation of data, (3) failure to use standard nomenclature, (4) second-hand opinions, and (5) unbridled romanticism.

All available data of reasonable reliability prove that (1) these popular legends are false, (2) the record for 1870-1947 shows a degree of relative regularity in behavior that is amazing for a geyser of such great size and complexity of function, (3) the amazing stability of the average interval between the initiations of eruptions, observed as 65 1 minutes in 1878 and as 65 12 minutes for the entire period 1870-1947, is unequalled, (4) it cannot be shown that since discovery in 1870 the behavior of the geyser has changed fundamentally in any way, although its idiosyncrasies have become better known, and (5) the same fallacious statements were being made in 1879. Old Faithful Geyser richly deserves its world-wide fame.

INTRODUCTION

OLD Faithful Geyser, the most celebrated geyser in the world, is in the Upper Geyser Basin in the valley of the Firehole River in the southwest-central part of Yellowstone National Park, Wyoming. Although discovery and naming of the geyser are credited to the Washburn-Langford-Doane expedition of 1870 because of its monumental work in making known the magnificent features of Yellowstone, the geyser was well known to James Bridger and other far-ranging "mountain men" for many years before 1870, and to the aborigines for untold years or centuries before that time. The origin of the geyser was discussed by Bauer and Marler.[3]

The writer became deeply interested in the geyser in 1932, and in addition to observations made as time permitted through 1939, he studied it daily from early June to mid-September

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[1] Published with permission of the Director, U. S National Park Service.

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[3] Bauer, C. M., and Marler, George. Old Faithful, an example of geyser development in Yellowstone Park. Assoc. Pacific Coast Geographers Year Book, 5, 1939, pp 45-48. Northwest Science, 13, 1939, pp. 50-55.

1937 while stationed at Old Faithful. This paper embodying results of the study before and during the time the writer was an employee of the U. S. National Park Service was prepared for publication in 1938, but the report of an extraordinarily long interval between two eruptions in that summer led him to postpone publication until additional perspective might be obtained from additional records. Observations for 1938, and by the U. S. National Park Service for 1939-1947, have been added, and all available data of acceptable accuracy therefore are correlated in this paper.

The writer is grateful to the following friends in the U. S. National Park Service for courtesies that facilitated his study: Dr. C. Max Bauer, Chief Geologist of the U. S. National Park Service; Chief Park Naturalist David DeL. Condon and Park Naturalist W. Verde Watson of Yellowstone National Park; and Carl Berryman, George Marler, Herbert T. Lystrup, W. Leon Evans, W. E. Kearns, Jennings King, Lloyd Sweetman, John Thompson, Julius Roller, and H. R. Woodward. He is grateful to J. E. Haynes, authorized Park Photographer, for the superb photographs illustrating this paper; for the record of his observations in 1910; and for many courtesies during the study.

No record of the behavior of Old Faithful Geyser would be complete without appropriate tribute to the painstaking work of A. C. Peale, whose records for parts of 1872 and 1878 and summary of the observations by other persons from 1870 to 1879[4] provide most of the available information concerning the early history of the geyser. The detail and the excellent form in which his records were kept have never been excelled. The writer found Peale's record and nomenclature so good that it was continued as part of the foundation for his own paper on geyser nomenclature.[5] Many garbled interpretations of geyser behavior have resulted from lack of appreciation and respect for the excellent prior nomenclature established by Peale, and to the lack of forethought and care.

The accurate determination of time seems to have received little attention from observers, who, by failing to standardize their chronometric instruments, have left us only the unfortunate record of geyser behavior as interpreted by the particular timepiece of unknown accuracy. Although facilities for accurate control of the writer's watch were not available in the summer of 1937, and it was not reset during the summer, the watch was slightly less than 3 minutes ahead at the end of the summer.

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[4] Peale, A. C Thermal Springs (of Yellowstone National Park): Old Faithful Geyser. U S Geol and Geog Survey of the Territories (Hayden), 12th Ann Rept, pt 2, 1878 (1883), pp. 220-229

[5] Fix, Philip F Nomenclature of geyser eruptions. Jour. Geology, 47, 1939, pp 99-104.

THE OLD FAITHFUL GEYSER OF FOLKLORE

It is commonly supposed that "in the good old days," Old Faithful Geyser erupted with the precision of clockwork. Many persons claim to have witnessed such behavior, but they have no *records* to show. Many of the old frontier men who were guides in the early years of the Park fostered such a belief as an integral part of the folklore of the Wild West then in vogue, and many visitors were so overwhelmed by the wonder of what they saw in Yellowstone that they failed completely to distinguish between fact and fiction when it was before their very eyes.

The writer has frequently heard it said that Old Faithful Geyser does not erupt at night, nor in the winter. One visitor said that on a previous visit several years before, she had seen Old Faithful erupting from another mound of sinter in the vicinity. Another said that she had seen the geyser erupt black smoke at the time of her previous visit 6 years before. The writer learned that what she actually saw was the water column of the geyser silhouetted against the early morning sun. The height of the geyser eruptions and the volume of water discharged likewise have been the subjects of unbridled imagination.

There is a fatal lack of conformity in the tales of what the geyser is supposed to have done. Some say that the eruptions occurred precisely every 60 minutes, but others claim with

PLATE 1

Fig. 1. This world-famous photograph shows an eruption of Old Faithful Geyser on a windless day with the water curtain falling around the foot of the geyser column (Haynes Inc. photo No. 13040).

Fig. 2. Eruption of Old Faithful Geyser on an average day silhouetted against the light, with a light wind carrying the curtain of water to fall beyond the geyser column (Haynes Inc. photo No. 10161).

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equal vehemence that the eruptions occurred precisely every 61, 63, 65, 66, or 67 minutes. Baseless rumors still are being manufactured. The writer heard on a radio program in 1938 that "Old Faithful Geyser erupts every 62 seconds." Furthermore, such precision is utterly impossible for a geyser of the magnitude and complexity of Old Faithful, whose eruptions are the result of the coordinated teamwork of a very large number of factors that differ greatly in kind and degree.

The fatal lack of conformity in the popular legends proves that they are unworthy of serious consideration. Examination of the written record of observations in 1878 alone extinguishes all such claims.[6]

THE OLD FAITHFUL GEYSER OF FACT

THE RECORD OF FREQUENCY

The feature of Old Faithful Geyser that has most intrigued the imagination of visitors in Yellowstone is the interval, which is that period of time between the initiations of two successive eruptions. In Old Faithful and other geysers of the same general type, a preliminary phase marked by the discharge of preliminary spurts of water precedes the initiation of the eruption phase proper.[7] The average interval for 1870-1947 is of greatest interest, but the length of each individual interval also is of considerable interest in geyser studies. The calculation of intervals in table 1 includes only those observations known to be reasonably accurate and whose method of calculation is known.

Table 1 does not show an interval on the night of July 13, 1938, that was reported to be 115½ minutes long, and that is generally thought to be an error that resulted from overlooking an eruption during that time. Such an interval approximately twice the length of the average interval, and far longer than any other known, occurring in the darkness of night inevitably is suspect because the writer and many others know, from deliberately awaiting eruptions of the geyser in broad daylight, how many times Old Faithful succeeds in attaining full eruption unnoticed when water-spray and sound are whisked away by the wind.

[6] Peale, A. C, op. cit, pp. 224-228.

[7] Fix, Philip F., op. cit., pp. 99-102.

TABLE 1

Interval between Initiations of Eruptions, Old Faithful Geyser, 1870-1947[a].

Year	Observer	Eruptions seen	Interval		
			Minimum	Maximum	Average
1870	Doane				50^{m}
1870	Washburn			.	60^{m}
1872	Peale	17	65^{m} 00^{s}	70^{m} 34^{s}	67^{m} 54^{s}
1873	Comstock	11	52^{m} 05^{s}	77^{m} 35^{s}	63^{m} 48^{s}
1875	Dana and Grinnell	24		...	65–66^{m}
1877	Sherman and Poe	7	62^{m}	80^{m}	67^{m}
1878	Peale	97	54^{m} 04^{s}	78^{m} 00^{s}	65^{m} 06^{s}
1879	Mitchell			...	70^{m}
1879	Seguin		..		45^{m}
1883	Hallock		...	..	65^{m}±
1928	Allen and Day	6	57^{m} 30^{s}	74^{m}	..
1928–9	Baker	38	58^{m} 20^{s}	75^{m}	..
1932	Robertson	1187	38^{m}	81^{m}	65 7^{m}
1937	Fix (July 17)	2		84^{m}	.
1937	Fix (Aug. 25–Sept. 3)	23	42^{m} 58^{s}	79^{m} 47^{s}	67^{m} 38^{s}
1938	Woodward[b]	771	...	...	66^{m} 30^{s}
1939	U.S.N.P S.	500	40^{m}	86^{m}	66 1^{m}
1940	U.S.N.P.S	1075	39^{m}	87^{m}	67^{m}
1941	U S.N P S.	1577	38^{m}	91^{m}	66.3^{m}
1942	U S.N.P S.	1452	41^{m}	90^{m}	60 05^{m}
1943	U.S N.P S	376	39^{m}	84 5^{m}	63 45^{m}
1945	U.S.N.P S.	940	39^{m}	85^{m}	63.77^{m}
1946	U.S N P S	1495	34^{m}	86^{m}	63 2^{m}
1947	U.S.N.P.S	1500	39^{m}	89^{m}	63.3^{m}
1870–1947		11098	34^{m}	91^{m}	65.12^{m}

(11048 used to calculate average interval)

[a] Data for 1870-1879 from A. C. Peale, op. cit., p. 229. Data for 1883-1932 from E T. Allen and Day, A. L., Hot Springs of the Yellowstone National Park, Carnegie Inst. Washington, Publ. 466, 1935, pp. 178, 184. Observations by Robertson in 1932 apparently are on basis used by the writer in this paper. Those by Allen and Day, and presumably Baker, were based on "the quiet intervals between eruptions," a method considered inaccurate by the writer because lengths of the preliminary phase and the duration vary considerably. The writer added 4½ minutes to each of their figures to convert them approximately to the basis used in this paper Data for 1939-1947 from personal communication by David DeL. Condon.

[b] Woodward, H. R Observations of Old Faithful Geyser U S National Park Service, Yellowstone Nature Notes, vol. 15, 1938. pp. 40-41.

The observed range between the maximum and minimum intervals, and the greatest departure from the average interval, for the entire period of record are shown in table 2.

The absolute minimum, maximum, and average intervals, and the absolute range of Old Faithful Geyser are unknown; tables 1 and 2 show only approximate values based upon a relatively small number of observations. It is surprising to find that in the 78 years of record, while the total number of eruptions observed had increased 350-fold, the observed range between maximum and minimum intervals was only doubled! It is also very remarkable that Peale observed an average interval of 65.1^{m} from 97 eruptions seen by him in 1878, and that 70 years later the average interval calculated from 11,048 eruptions is 65.12^{m}!!

TABLE 2

Observed Range between Maximum and Minimum Intervals, and Greatest Departure from the Average Interval, Old Faithful Geyser, 1870-1947

Year	No. years observed	Eruptions observed	Observed range	Greatest departure from 1870-1947 average	
				Average - minimum	Maximum - average
1870	1[a]	?	10^{m}+	15^{m} 07^{s}	?
1873	3	40±	27^{m} 35^{s}	15^{m} 07^{s}	12^{m} 28^{s}
1879	10	160	35^{m}	20^{m} 07^{s}	14^{m} 53^{s}
1937	68	1391	46^{m}	27^{m} 07^{s}	18^{m} 53^{s}
1947	78	11048	57^{m}	31^{m} 07^{s}	25^{m} 53^{s}

[a] The total number of eruptions, and the minima and maxima observed by Doane and Washburn are not known, but are greater than the difference of 10^{m} between the average interval given by each.

THE RECORD OF HEIGHT

The measurement of height of geyser eruptions is of great interest. Their height is exceedingly difficult to estimate, and very greatly misjudged. Most persons cannot estimate accurately a distance laid out on the ground under the most familiar circumstances; much less are they able to estimate a vertical distance against the intangible blue sky. Not only is the average person unable to estimate the apparent height, but he has little understanding of the conversion of apparent into true height. It is surprising how many persons completely lack the mathematical basis for understanding an instrumental determination of height by vertical angle, yet many of these persons assert their ideas of geyser height as absolute fact. These erroneous ideas have had wide circulation, whereas the scientific record has not. The legend arose that Old Faithful Geyser no longer erupts to the great height of former years.

Inasmuch as very experienced observers are unable to estimate the exact height, the only useful data are actual instrumental measurements of height in which the method used is stated clearly. Table 3 shows the acceptable available records of heights actually measured. The earlier measurements of height are a little low in comparison with the later measurements because of a change in method used. Peale measured geyser height to the top of the unbroken column of water, and apparently the other early observers did likewise. The 1910 measurements by Haynes were "the height of the water column."[8] The heights given by Allen and Day, Baker, and the writer are known to have been to the top of the highest splashes, a more accurate method.

TABLE 3

Measured Height of Eruptions, Old Faithful Geyser, 1870-1947[a].

Year	Observer	Eruptions measured	Height of eruption (feet) Mimimum	Maximum	Average
1871	Barlow	?	?	138	?
1872	Peale	5	106	130	122
1875	Dana and Grinnell	24	?	?	115
1878	Peale[b]	11	106	139	125
1910	Haynes[c]	5±	?	?	150
1928	Allen and Day[d]	6	114	150	133
1928–9	Baker[d]	38	110	160	?
1937	Fix[e]	19	111	184	146
1871–1878		40+	106	139	119
1928–1937		63[f]	110	184	143
1870–1947 summary		108+[g]	106	184	130[g]

[a] Unless otherwise specified, data in this table are from Peale, op. cit., pp. 221-229. Peale used a gradienter, and his 1872 baseline was 100 feet long.

[b] Measurement by Ladd in 1878, considered inaccurate by Peale, are not included in this table.

[c] Haynes, J. E., personal communication

[d] Allen, E. T., and Day, A. L., op. cit., p. 185. Measured with Klausner hypsometer. Baseline 250 (?) feet long, measured with steel tape

[e] Measured with Brunton compass and 660 foot baseline measured with steel tape. Vertical angles believed very accurate within limitations of the instrument, as result of writer's long experience.

[f] Average calculated only from the 25 measurements for which averages are stated.

[g] Average is lowered by inclusion of older measurements made to top of water column only. Measurements for 1928-1937 made to top of highest splash of water. Average is calculated only from the 70± measurements for which the averages are given.

Haynes, J E., personal communication

Many observers seem to have overlooked the fact that the wind decapitates the water column of a geyser by whipping away the upper part as spray. The writer's observations have led him to believe that more than 50 feet may at times be subtracted from the column of Old Faithful Geyser by the wind, but neither maximum nor average values can be estimated because of inadequate data. Variations in the height of eruptions resulting from the complexity of geysers are imperfectly understood, and they cannot be evaluated without adequate records of wind influence upon the height and operation of the geyser. The observations made by the writer in 1937 include a record of the wind.

THE RECORD OF WATER DISCHARGE

Estimates of the volume of water discharged by Old Faithful Geyser have been most romantically colored. The value generally supposed in the past seems to have been about 750,000 gallons of water per eruption. This volume, necessarily being discharged in little more than 4 minutes at a diminishing rate, means a volume of about 450 second feet. This volume is 3½ times that of the Firehole River two miles downstream from Old Faithful Geyser![9] One glance at the runoff streams from the geyser eruption shows that such volume is impossible. The reliable figure of 10,000 to 12,000 gallons per eruption, a very large volume for the circumstances, was measured in 1929 by Boyer at the request of Allen and Day.[10]

COMPARATIVE REGULARITY OF OLD FAITHFUL GEYSER

The relative regularity of Old Faithful Geyser is so astonishingly great for the period 1870-1947 that the world-wide fame accorded the geyser is deserved beyond equivocation, but the common belief that Old Faithful is the *only* regular large geyser in Yellowstone is incorrect. The degree of the regularity of several other large geysers over long periods of time is not known accurately because their lesser fame has resulted in even fewer written records than for Old Faithful. The Riverside Geyser in the summer of 1937 showed a regularity greater than that of Old Faithful.[11] Valentine Geyser [12] in the summer of 1936 erupted with a regularity comparable with that of Old Faithful, but there have been long periods of time before and since when it was irregular or inactive because of interference from its neighbors. Daisy Geyser and several others are known to have been relatively regular over long periods also.

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[9] Allen, E. T., and Day, A. L., op. cit., p. 269.

[10] Allen, E. T, and Day, A. L., op. cit., p. 184.

SUMMARY

The known acceptable records of eruptions of Old Faithful Geyser from 1870 to 1947 present the following conclusions:

1) Each observation is a special case. The *absolute* minimum, maximum, and average intervals between initiations of eruptions, and the *absolute* range between the maximum and minimum intervals are unknown. The approximations to 1947 are as follows·

Average interval	65.12^{m}
Minimum interval	34^{m}
Maximum interval	91^{m}
Range (Maximum — minimum)	57^{m}

2) The *average interval* of Old Faithful Geyser is amazingly stable. Peale found, from the 97 observations made in 1878 that constitute the only series of observations "in the good old days" comprehensive enough to afford a reliable average, that the average interval was 65.1^{m}. *Seventy years later*, the average interval calculated from 11,048 eruptions is 65.12^{m}!!

3) No change in average height of eruptions can be shown for the period of record, 1870-1947, although the records for recent years show greater heights than formerly, the opposite of the popular legends. Exact comparison of measured heights from 1870 to 1947 is impossible because of the change in method (formerly to top of the unbroken column of water, now to the top of the highest splash of water) that makes the earlier records seem a little low.

4) Old Faithful Geyser discharges about 10,000 to 12,000

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[11] Fix, Philip F. The phenomenal regularity of Riverside Geyser U S National Park Service, Yellowstone Nature Notes, 17, 1940, pp. 21-22.

[12] Fix, Philip F. Valentine Geyser. Idem, 14, 1937, pp. 31-37.

gallons of water per eruption, a very large volume for the circumstances.

5) It is surprising to find that in the 78 years of record, while the total number of eruptions observed increased 350 times, the observed range between maximum and minimum intervals was merely doubled.

6) The description by Peale of an eruption of Old Faithful Geyser in 1872 or 1878 might well have been written in 1947.

7) It cannot be shown that since the discovery in 1870 the behavior of Old Faithful Geyser has changed fundamentally in any way whatsoever, although its idiosyncrasies have become better known. The evidence points conclusively to the opinion that no fundamental change occurred in the geyser from 1870 to 1947, nor probably for a very long time before 1870.

8) Future research will reveal factors, trends, and cycles at present unknown because of inadequate records. Future records should include data on weather conditions and kindred factors that unquestionably play a part in eruptions of Old Faithful Geyser.

9) Old Faithful Geyser is not the only large, relatively regular geyser in the world, but to point out that there are other large but less well known geysers that have been relatively regular for short periods of time does not detract from the world-wide fame that Old Faithful Geyser so richly deserves because of its amazing degree of regularity for so many years.

10) It is significant to note, as did Peale, that as early as 1879, if not earlier, the same fallacious statements were being made about Old Faithful Geyser as are being made today. In his report for 1879, Superintendent P. W. Norris said: "Old Faithful, though still the favorite geyser, has certainly lost prestige by an occasional prolongation of its intervals between eruptions of from 5 to 15 or 20 minutes during . . . the past season."[13] It need only be pointed out that as early as 1873, six years before the Norris report and only three years after the official discovery of the geyser, a range of 27 minutes 35 seconds had been observed. It is clear that no further comment is needed concerning the amazing regularity of Old Faithful Geyser since 1870!

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[13] Peale, A C., op. cit., p 229.

1948 ACTIVITY

The remarkable regularity of Old Faithful Geyser is still further proved by the following data for 1948 that became available after this paper had been set in type.[14]

Number of eruptions observed	1,486
Average interval	64.1^{m}
Minimum interval	33^{m}
Maximum interval	85^{m}

The following summary correlates these with earlier records; the insignificant changes resulting from 1,486 additional observations is noteworthy:

Average interval, 1870-1948 (based on 12,534 eruptions)	65.01^{m}
" " 1870-1947 (based on 11,048 eruptions)	65.12^{m}
Minimum interval, 1870-1948	33^{m}
" " 1870-1947	34^{m}
Maximum interval, 1870-1948	91^{m}
" " 1870-1947	91^{m}
Range (maximum — minimum), 1870-1948	58^{m}
" " " 1870-1947	57^{m}

U S. GEOLOGICAL SURVEY
LOGAN, UTAH

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[14] Marler, George. Geyser activity in the Old Faithful area—1948. U S. National Park Service, Yellowstone Nature Notes, 22, 1948, p. 64.

RADIOACTIVITY OF OCEAN SEDIMENTS. VI. CONCENTRATIONS OF THE RADIOELEMENTS IN MARINE SEDIMENTS OF THE SOUTHERN HEMISPHERE

WILLIAM D. URRY

ABSTRACT. It has been reported in previous publications of this series that the mode of variation of the radium concentration below the ocean bottom affords a method of determining time in ocean sediments. Hitherto, these researches were confined to the northern hemisphere Similar studies in the southern hemisphere, combined with the necessary geological and biological investigations, should provide an answer to the question of the contemporaneity of glaciation in the northern and southern hemispheres. Measurements of the radium content as a function of depth in the sediment are presented here for ocean-bottom cores secured by the U.S. Navy Antarctic Expedition of 1946-47.

INTRODUCTION

MANY measurements of the radium content of deep-ocean sediments have been reported in previous publications of this series.[1-5] A number of determinations of the radium content of "grab samples" of bottom sediments were given by Piggot.[1] Among these were measurements on samples secured in the southern hemisphere, but until the present investigation was undertaken the variation of the radium content below the surface of the ocean bottom south of the equator had not been studied.

One might inquire why interest should be evinced in comparing the radioactive relations in the deposits of the southern hemisphere with those in the sediments of the northern hemisphere. Variation of the radium content has been studied in detail for a number of cores secured in the northern hemisphere;[6] and there is no reason to believe that the overall relations between the members of the uranium-238 series are simply dependent on the latitude and longitude. The present results fully bear out the universal nature of these radioactive relations in all marine sediments with the possible exception of inshore deposits.

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[1] Piggot, C. S, 1933. The Radium Content of Ocean-Bottom Sediments. AM. JOUR. SCI, 25, 229.

[2] Piggot, C. S, and Urry, Wm. D., 1939 The Radium Content of an Ocean-Bottom Core J. Wash. Acad. Sci., 29, 405

[3] Piggot, C. S., and Urry, Wm. D, 1941 Radioactivity of Ocean Sediments. III Radioactive Relations in Ocean Water and Bottom Sediments AM. JOUR SCI, 239, 81.

[4] Piggot, C. S., and Urry, Wm D, 1942. Radioactivity of Ocean Sediments IV The Radium Content of Sediments of the Cayman Trough. AM. JOUR. SCI, 240, 1

[5] Urry, Wm. D, and Piggot, C. S., 1942 Radioactivity of Ocean Sediments V. Concentrations of the Radioelements and their Significance in Red Clay. AM JOUR SCI, 240, 93

Our main interest in studying many core-samples of deep-water sediments from all the oceans arises from the fact that the lack of radioactive equilibrium in the sediments, at the moment that they are deposited, provides a method of measureing time below the ocean floor for the order of the last half-million years.[7]

The concentration of uranium in the material being deposited on the ocean floor is insufficient to support its ionium content, and, therefore, the excess ionium decreases—by 50 per cent of the remainder every 82,000 years. For practical purposes, the excess ionium disappears in 400 to 500 thousand years. Thereafter, the ionium content is constant because the uranium content can be considered inexhaustible for periods of millions of years.

No reliable method is yet available for the rapid measurement of the minute amounts of ionium scattered universally throughout the Earth's crust, but fortunately the readily determined radium suffices to provide the necessary data concerning its parent ionium. With one exception[5] (which in the light of recent results may not actually be an exception), the material depositing on the ocean floor contains less radium than the ionium can support. Consequently, the following general picture for the variation of the radium content with time, and therefore, with the depth of the buried sediment, is obtained. The radium content increases below the surface rather rapidly, unless the rate of deposition is phenomenally high, until it is in equilibrium with the ionium content. Below the depth representing the time interval necessary to establish this equilibrium the radium content diminishes only so fast as the ionium content decreases. Finally, if the material at the bottom of the core be considerably older than 500,000 years, the curve of radium content against depth should exhibit a flat tail and the constant radium content is then a measure of the uranium content as well as of the ionium content, all the elements in the uranium-238 series being in radioactive equilibrium. This general picture of the variation of the radium content with the depth of the sediment is found to be true, not for a few specific examples, but for all the cores of deep-ocean sediments that have been examined. Shelf deposits remain to be investigated. In a single long core of inshore sediment in the present suite, the radium content remained extremely constant from top to bottom in agreement with the results of Sanderman and Utterback[8] working with very short cores collected in the north Pacific.

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[6] Piggot, C. S., and Urry, Wm. D., 1941. Time Relations in Ocean Sediments. Bull. Geol. Soc. Amer. 53, 1187

[7] Urry, Wm. D, 1942 The Radioelements in Non-equilibrium Systems. Am. Jour. Sci., 240, 426

All the above phenomena associated with the variation of radium content with time beneath the ocean floor can be given a complete quantitative description with the equations governing the growth and decay of the radioelements.[7] In this manner it is possible to analyze the experimental curve of radium content as a function of depth and to assign a date to the sediment at any desired depth.

In most cores of deep-sea sediments there is a wealth of evidence of the changing geological, geophysical, climatological, and oceanographical conditions of the immediate past. Studies of the lithology and biology of cores of marine sediments have produced a record of the history of these changing conditions and this record appears to be preserved in a more complete and less disturbed state in many ways, than that revealed by studies of the continents.[9] For a satisfactory solution to many of the problems of the history of the immediate past it is essential to establish a basis for correlating the evidence for specific events, such evidence often being uncovered at widely separate localities, in very different types of sediments, and in deposits accumulating at very different rates. An outstanding example of this type of problem is the question of contemporaneity of glaciation in the northern and southern hemispheres. It was with a view to providing an answer to this question in the near future that the present work was undertaken. Obviously, there is only one sound basis for

[8] Sanderman, L A, and Utterback, C. L, 1941. Radium Content of Ocean Bottom Sediments from the Arctic Ocean, etc. Sears Foundation: J. Mar. Res. 4, 132.

[9] Bradley, W. H, et al., 1940. Geology and Biology of North Atlantic Deep-Sea Cores. U S.G.S. Prof. Paper 196-A.

the correlation of the evidence; it must be dated with respect to a fixed time, preferably the present.

Presentation of the time-depth analyses of the curves of radium content against depth for these cores from the southern hemisphere would serve no useful purpose here because the results of studies of the lithology, of the foraminifera, and of other properties of these sediments are not yet available. In view of the fact that the accumulation of these necessary data may require appreciable time, it seemed advisable to report here the radioactive measurements that will eventually form the basis for dating these sediments. Furthermore, a knowledge of the concentrations of the radioelements in deep-sea sediments of all the oceans is necessary to understand the relations existing between the radioelements in ocean water and in bottom sediments. These new results will form an important addition to the body of data necessary for a consideration of the principle of dynamic equilibrium in the ocean as applied to the radioelements. Attention was drawn to the complementary nature of the Ra/U ratios in water and sediment some years ago[3]; but the transiency of the radioelements complicates the problem, and little could be accomplished with the data in existence at that time. This problem will be taken up elsewhere following publication of new determinations of the radium and uranium content of river and ocean waters. Studies of this type are of interest because they promise to provide an insight into the operation of such mechanisms as co-precipitation, selective adsorption, and concentration in organic matter.

SOURCE OF THE MATERIAL

Dr. J. L. Hough, formerly of Woods Hole Oceanographic Institution,[10] accompanied the U. S. Navy Antarctic Expedition of 1946-47 on board the U.S.C.G.C. *Northwind*. He was successful in securing ten core samples of the ocean bottom varying in length from six inches to eight and one-half feet. Six of these cores, that is, all of those exceeding five feet in length, were chosen for determination of the radium content. The cores had been sealed and preserved in Lucite liners following withdrawal of the liners from the coring tube. The liners were split longitudinally and one-half was sampled at appropriate intervals of depth. Full details concerning the apparatus employed to secure these samples, and the procedures followed in treating the samples, will undoubtedly be described by a competent authority elsewhere.

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[10] Present address: Geology Department, University of Illinois, Urbana, Illinois

An ambitious program of many studies was planned for these cores, and, therefore, a minimum amount of material could be spared for any given part of the program. There is a further limitation in the amount of material available because it is highly desirable to work with samples that represent no more than a depth of one cm. in the sediment. The quantities employed for the radioactive measurements varied from as little as 0.3 gram near the top to 2 to 3 grams near the bottom. These amounts proved to be just sufficient to measure the radium content with the required accuracy.

The titles used to describe the sediments in this paper are tentative. They are based on preliminary observations, chiefly of color and texture, and they certainly fail to describe all the less obvious changes in the type of sediment that occur within a given core, particularly in the case of those classed as glacial marine.

The locations of the sediments sampled by these cores are shown on the charts in figure 1.

EXPERIMENTAL

The apparatus and procedure for the determination of minute quantities of radium have been fully described elsewhere.[11] Most specimens of ocean sediments lose radon on standing. The determination involves the measurement of the gaseous radon in equilibrium with radium, and, therefore, in order to insure that all the radon was collected in the ionization chamber the present samples were heated as previously described[11] for 40 to 60 minutes at 550°C in a stream of nitrogen and sealed for at least 30 days prior to measurement. At the time of measurement the gas in the storage tube was collected and added to that derived from fusing the sample. The weight of sample necessary for expressing the result was determined after drying the specimens at 110°C for 24 hours and prior to heating to 550°C. Heating to the

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[11] Urry, Wm. D, and Piggot, C. S., 1941. Apparatus for Determination of Small Quantities of Radium. Am. Jour. Sci., 239, 633.

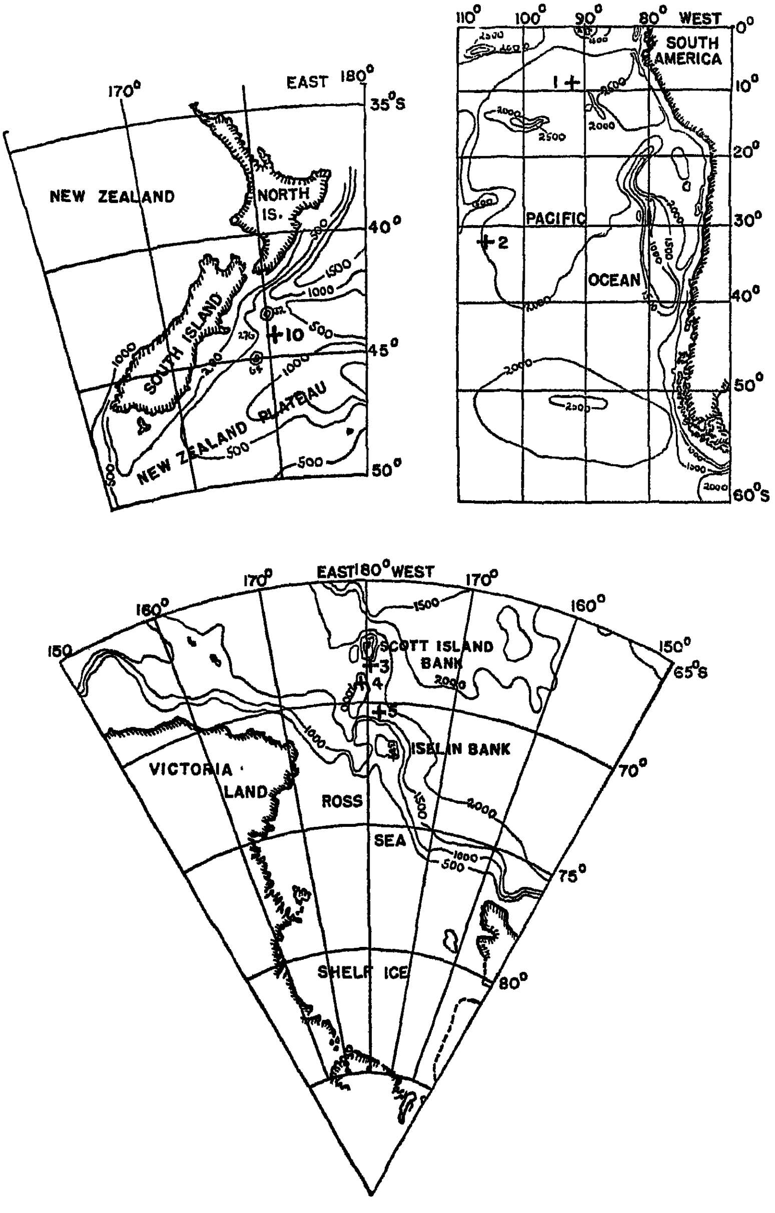

Figure 1. Partial tracings of charts of the U. S. Navy Hydrographic Office showing the location of the sites where core samples of ocean-bottom sediments were secured by the U. S. Navy Antarctic Expedition of 1946-1947. Only those cores studied for radioactivity are indicated. All depths are given in fathoms.

higher temperature avoids certain difficulties otherwise encountered in fusing the sample to liberate the radon.[11]

The root mean square probable error, insofar as the statistical fluctuations of the ion current and errors in measuring the records and setting calibration voltages are concerned, is about 1.5 per cent for most of the determinations. In a few cases where only very small samples were available the probable error was 3 per cent. Periods of observation varied from 5 to 20 hours depending upon the total amount of radium to be measured.

The absolute accuracy of the determinations is another matter. For the determination of time in ocean sediments it is only necessary to know the relative values of the radium content in any given core. However, for other purposes, such as have been indicated above, the absolute values are of interest. Radium solutions containing 10×10^{-12} g. radium, prepared by the National Bureau of Standards, were employed to calibrate the apparatus. From 1941 to 1948 repeated calibrations have always yielded the same calibration constant to within 2 per cent. This, however, affords no complete check on the agreement between various laboratories and various methods, and to provide for such checks the National Bureau of Standards prepared in 1939 a suite of rocks to serve as a means of intercalibration. With the exception of a few inhomogeneous granites, the agreement between the radium determinations at the Massachusetts Institute of Technology, the National Bureau of Standards, and the Geophysical Laboratory was satisfactory and the standard rock samples have been certified to have a given radium content within appropriate limits.[12]

In order to intercheck the apparatus used for the ocean sediments, which measures the total ion current, with another apparatus in use at the Geophysical Laboratory which determines radium by counting alpha particles,[13] one of the suite of rock samples used as standards of radioactivity has been measured at frequent intervals with the results shown in Table I.

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[12] National Research Council, 1938- . Minutes of the Committee on Standards of Radioactivity R. D. Evans, Chairman. (The work of this committee is now the responsibility of the Sub-committee on Radiogeology of the N.R.C. Committee on Nuclear Science.)

[13] Davis, G. L., 1947 Radium Content of Ultramafic Igneous Rocks. I. Laboratory Investigation. Am. Jour. Sci., 245, 677.

TABLE I

Standard Rock Kimberlite in 10^{-12} g. radium per g
Measurements of the Radium Content of the

NBS value		0.59 ± 04	some sealed, some not sealed
Geophysical Laboratory			
ion current	1940	0.64 ± .01	sealed
	1942	0.65 .02	sealed
	1947	0 64 02	not sealed
counting	1941	0.67 ± .01	not sealed
	1941	0.64 01	not sealed
	1941	0.68 .01	sealed

On the basis of the measurements of the kimberlite, it might appear that determinations by both methods at the Geophysical Laboratory are slightly too high, but this is not the case when one considers the results from the three laboratories for all the standard rocks. The kimberlite has a very convenient radium content but is not to be highly recommended as the most suitable of the standards because of a tendency to spatter in the fusion furnace whether pre-heated or not. Had no allowance been made for the amount of spattered material, the values in Table I would have been in good agreement with the value assigned by the National Bureau of Standards. It can be stated with assurance that the radium contents of the sediments reported here are accurate to 10 per cent on an absolute basis.

In the analysis of the radioactive data for the purpose of finding the time-depth relations, it is important in some cases to express the radium content on a volume basis. This can be done with a knowledge of the original water content. Those samples in which water was to be determined were weighed within thirty minutes after opening a core liner and later heated at 110 to 120°C and re-weighed. It appears from the results in Table II that the ends of the core liner were not always well sealed. However, there was little evidence of shrinkage, indicating appreciable loss of water, except at the ends of one or two cores.

TABLE II

The Radium Content of Ocean-Bottom Cores;
U.S.C.G.C. *Northwind,* U.S. Navy Antarctic Expedition 1946-47

(Collected by Dr. J. L. Hough)

Core N-1, Lat 08°56′S, Long 92°05′W, 2150 fathoms, alternating red clay and globigerina ooze.

Specimen	Depth from top of the core (cm.)		Water content per cent by weight	Radium content in 10^{-12} g. per g.
		(red clay)		
N-1- 0	0.0 to	0.3	54.1 (65.5)[a]	18.95
N-1-10	2.0	2.5	—	19.81
N-1-11	4.0	4.5	—	22.39
N-1-14	12.0	12.5	—	28.07
N-1- 4	15.0	15.5	65.9	29.11
N-1-18	25.5	26.0	—	29.67
N-1- 5	30.0	30.8	63.9	26.36
N-1- 7	192.7	194.2	71.5	0.53
		(mixture — predominantly red clay)		
N-1- 2	6.0	6.5	57.3	20.38
N-1-13	9.0	9.5	—	20.43
N-1-17	22.0	23.0	—	20.38
		(mixture — about equal amounts of each type)		
N-1-12	6.6	7.2	—	16.90
N-1-15	18.0	18.5	—	15.86
		(Globigerina ooze)		
N-1-16	20.5	21.5	—	7.01
N-1-19	28.0	29.0	—	6.45
N-1- 3	32.5	33.0	64.3	6.18
N-1-21	40.5	41.5	—	4.75
N-1-23	59.0	60.0	—	4.71
N-1-24	80.0	81.0	—	2.85
N-1-25	100.0	102.0	—	1.40
N-1- 6	124.5	125.7	51.6	0.57
N-1-27	166.5	167.5	—	0.61

[a] Value corrected for observed shrinkage arising from loss of water while in storage in the Lucite liner.

TABLE II (Continued)

Core N-2, Lat. 32°21′S, Long. 105°55′W 1980 fathoms.
Red clay

Specimen	Depth from top of the core (cm.)		Water content per cent by weight	Radium content in 10^{-12} g. per g.
N-2- 0	0.0 to	0.3	45.8	9.08
N-2- 1	2.0	2.5	46.4	11.09
N-2- 2	6.0	6.5	47.0	13.19
N-2- 3	11.8	12.5	47.0	12.33
N-2-11	16.0	16.5	—	11.16
N-2- 4	25.0	26.0	50.1	7.30
N-2-14	50.0	51.0	—	2.35
N-2-16	80.0	81.0	—	0.72
N-2- 5	100.0	101.0	43.4	0.76
N-2- 6	137.3	138.3	41.5	0.69

Core N-3, Lat. 68°26′S, Long 179°35′W. 1800± fathoms.
Glacial marine

Specimen	Depth from top of the core (cm.)		Water content per cent by weight	Radium content in 10^{-12} g. per g.
N-3- 0	0.0 to	1.0	—	3.62
N-3- 1	2.0	2.5	38.6	3.49
N-3- 2	8.0	8.5	44.7	4.30
N-3-10	15.0	16.0	—	5.68
N-3- 3	24.0	24.5	47.0	5.07
N-3-12	33.0	34.0	—	4.85
N-3- 4	50.0	51.0	48.4	3.92
N-3-15	80.0	81.0	—	2.41
N-3- 5	100.0	101.0	47.0	1.77
N-3-16	125.0	126.0	—	0.97
N-3- 6	169.0	170.0	47.5	0.62
N-3- 7	227.0	228.0	41.6	0.62

Core N-4, Lat. 69°12′S, Long 179°34′E. 2040 fathoms.
Glacial marine

Specimen	Depth from top of the core (cm.)		Radium content in 10^{-12} g. per g.
N-4- 0	0.0 to	0.5	2.62
N-4- 3	10.0	10.5	3.37
N-4- 5	20.0	21.0	3.70
N-4- 6	30.0	31.0	4.08
N-4- 7	49.8	51.0	4.26
N-4- 8	80.0	81.0	2.65
N-4- 9	120.0	121.0	3.46
N-4-10	160.0	161.0	2.21
N-4-12	225.0	226.0	1.64
N-4-13	259.5	260.5	1.53

No measurements of water content were made in this core

TABLE II (Continued)

Core N-5, Lat 70°17′S, Long. 178°23′W. 1635 fathoms.
Glacial marine

Specimen	Depth from top of the core (cm.)		Water content per cent by weight	Radium content in 10^{-12} g. per g.
N-5- 0	0.0 to	1 0	32 1	1.70
N-5- 1	4.0	4.7	40 2	1.79
N-5- 2	9.0	9.6	46.8	2.13
N-5- 3	16 0	17 0	44 2	2.67
N-5- 4	25.0	26.1	43 4	3.03
N-5- 5	55.0	56 0	43 4	2.58
N-5-16	105 0	106 1	—	0.95
N-5- 6	130.0	131.0	45.7	1.29
N-5-17	165.0	166.1	—	1.12
N-5- 7	200 0	201 1	36.6	1.01
N-5- 8	241 3	242 5	40 2	0.77

Core N-10, Lat. 44°14′S, Long. 175°15′E. 300 fathoms.
Greenish gray mud

Specimen	Depth from top of the core (cm.)		Water content per cent by weight	Radium content in 10^{-12} g. per g.
N-10-0	0.0 to	1.0	28.9	1.17
N-10-4	50.0	51.0	35.0	1.14
N-10-6	100.0	101.0	32.5	1.18
N-10-7	140.0	141 0	33.3	1.11
N-10-8	175 0	176.0	29.3	1 18

RESULTS

All determinations of the radium content of the cores from the southern hemisphere are to be found in Table II. The variation of the radium content with the depth in each core is shown in figures 2, 3, and 4. The number of determinations necessary to define a curve varies appreciably for several reasons. In general, the samples must be more closely spaced near the top where the radium content is changing most rapidly, but extreme variations in the type of sediment and sudden changes in the rate of deposition will also require a more detailed survey than would otherwise be the case.

With the exception of the inshore deposit (core N-10), the maximum radium content in all deep-sea cores is to be found at some distance below the surface of the ocean bottom. One can gain a general idea of the recent rate of deposition from the depth at which such maxima occur. The time required

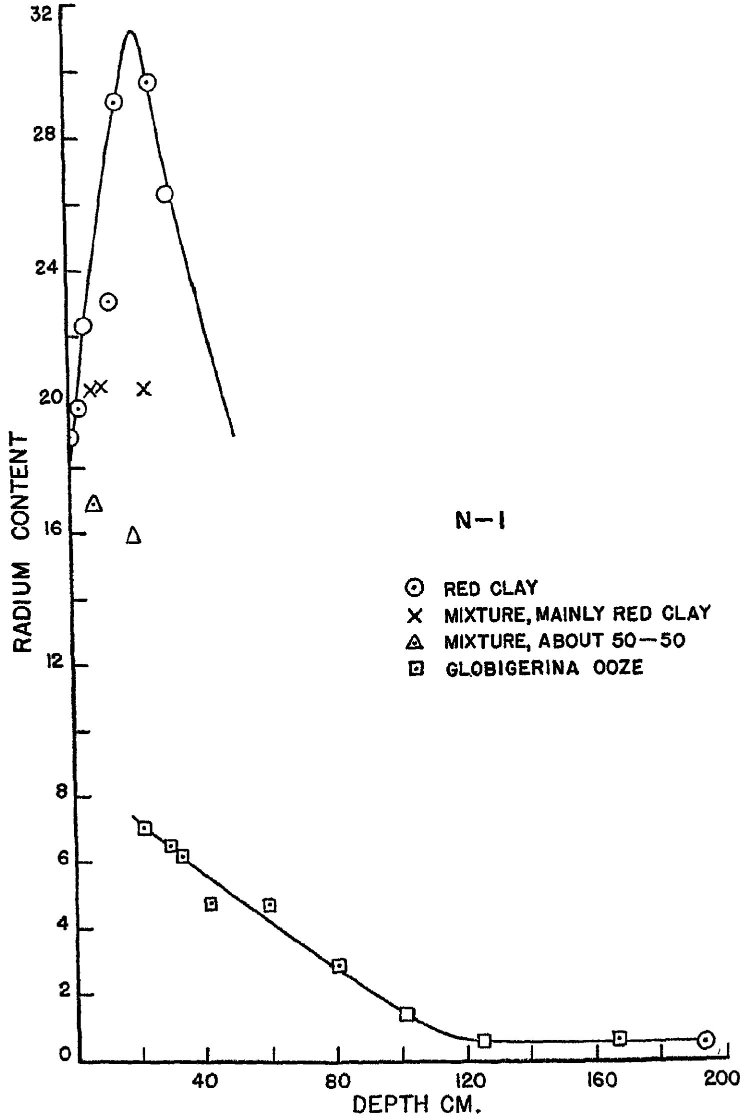

Figure 2. The variation of radium content with depth in a core of alternating red clay and globigerina ooze The radium content is given in units of 10^{-12} g. radium per g. of sediment dried at 110°C

to establish the maximum radium content is a function of the initial relation between radium and ionium, but it is not greatly dependent on this parameter and will be between 7000 and 9000 years.[7] A second obvious point on the time scale is where the radium content becomes independent of the depth in the core. This will occur in between 400 and 500 thousand years. Thereafter, this method of dating becomes inapplicable, but reasonable dates can be assigned below such depths if one assumes that the rate of deposition prior to 500 thousand years ago was not greatly different from the rate 400 thousand years ago, and if the distortion in securing the core sample is taken into account.[14]

INDIVIDUAL CORES

Core N-1. (*fig. 2*) Red clay and globigerina ooze have been alternately deposited at this site. It was hoped that two curves could be obtained, one for each type of sediment, and that these curves could be independently analyzed for time. If good agreement resulted, such a procedure would offer strong evidence in support of the basic principles underlying the measurement of time in ocean sediments. Unfortunately, red clay predominates in the top 30 cms., and insufficient globigerina ooze of the required purity is to be found above 20 cms. Below 30 cms., red clay of sufficient purity is unavailable except right at the bottom of the core. Thus, it is only possible to make the desired comparison between the depths of 20 and 30 cms.

It is interesting to note that the radium content, following establishment of complete equilibrium between radium, ionium, and uranium, is independent of the type of sediment. This means that the uranium content, which does not change with time, is the same for both types of sediment in this core and that the striking differences near the top and particularly in the middle are due to very different initial ionium contents.

The maximum recorded radium content in this core occurs at 25.8 cms. (n-1-18). There are very few samples of red clay that show as high a radium content as 29.67×10^{-12} g.

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[14] Piggot, C. S, 1941 Factors Involved in Submarine Core Sampling Bull. Geol. Soc Amer 52, 1513.

per g. and there is not one from the "Carnegie" collection of grab samples.[15]

This core sample must be representative of the sediment of more than half of all Pleistocene time at this location.

Core N-2. (*fig. 3*) This core sample, according to color and texture, is a red clay of very uniform character from top to bottom. Consequently, it is to be expected that the variation of the radium content can be expressed by a very smooth

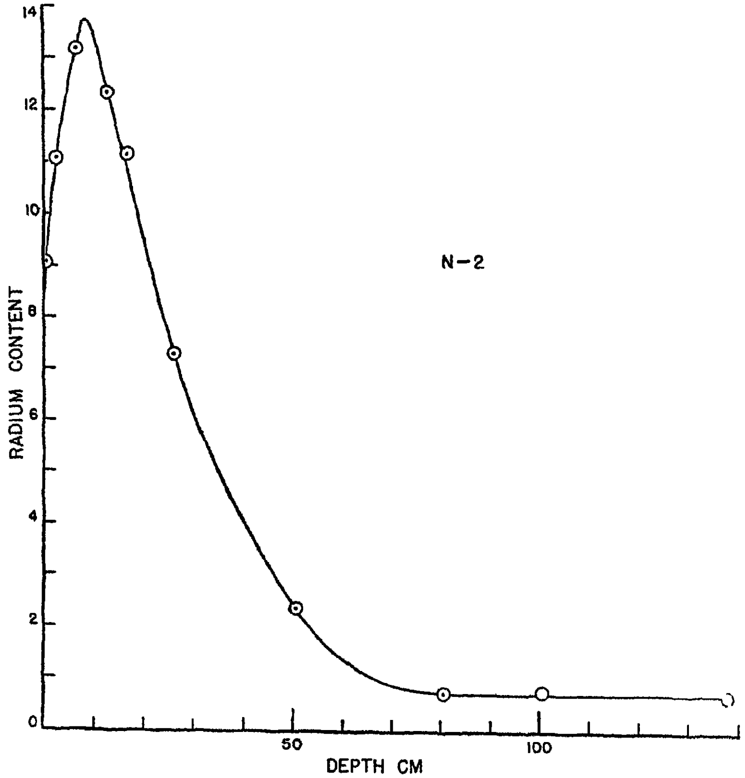

Figure 3. The variation of radium content with depth in a core of red clay. The radium content is given in units of 10^{-12} g radium per g. of sediment dried at 110°C.

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[15] With the aid of figures 2 and 3 it can readily be realized that the radium content of a grab sample has little significance because it depends upon a number of arbitrary factors—the depth of penetration of the grab, and, therefore, the kind of grab employed; whether or not the sample was homogenized; and, if not, the location of the sub-sample for the radium measurement.

curve. This curve exhibits the maximum below the top characteristic of all the other cores. Prior to these determinations, measurements in red clay had been made for only one core of any significant length.[5] The samples for these earlier determinations were undesirably long; and, hence, it was difficult to establish the shape of the curve near the top. For various reasons it was concluded that there was no maximum radium content below the surface; and, thus, the single curve for red clay constituted an exception to the general rule. In view of the curve obtained for the present core of red clay and for the red-clay portion of core N-1 in figure 2, this previous exception may be in doubt.

Core N-2 must be representative of the sediment of most of Pleistocene time.

Cores N-3, N-4, and N-5, (fig. 4) These cores were all secured from the relatively small area lying between Scott Island Bank and Iselin Bank in the Ross Sea, as shown in figure 1. Upon superficial examination they all look quite similar, but at certain depths the sediment is completely free of pebbles and small angular fragments and at other depths such glacial marine debris is abundant. The sediments of all three are tentatively classified as glacial marine. Bearing in mind the two particularly significant positions on the time scale—the maximum value of the radium content and the position where the radium content reaches an essentially constant value—a glance at figure 4 shows that the rate of deposition of the sediment was very different at the site of each one of these cores. It would be interesting to look for the reason for this behavior in continuous ocean-bottom profiles of the area if these were available. The sediment is deposited most rapidly at the site of core N-4 in a small trough south of Scott Island Bank. Events that are characteristic of a given time should have left their records at very different depths in these three cores.

A pronounced dip in the curve is noticeable at the bottom of core N-5. This is to be expected[8] in all cases where the sediment at the bottom is less than 400 thousand years old because of the distortion occurring when the sediment is cored.[14] The distortion causes a shortening of the core sample compared to the original sediment, the amount of the shortening being negligible near the top but increasing rapidly near the bottom. A similar effect is present in the curve for core N-4, but it is less noticeable in this region of the curve. Naturally, no such dip is to be seen when the constant equilibrium value of the radium content is attained at some distance above the bottom of a core. The distortion will have no influence on the dating of a specific event, but it is a factor of no slight significance when rates of deposition are to be determined.[6]

One point for each of the cores N-4 and N-5 lies considerably below the curve. Insofar as was possible, pebbles and fragments of rock were removed from the sample whenever they were found in the course of reducing the sample to powder. Such material was considered foreign to the sediment carrying the pertinent radium but some coarse detrital material may have escaped detection. Since such material is almost certainly of much lower radium content than that indicated by the curve, particularly if basic in character, the discrepancy is in the right direction.

Core N-10. (*fig. 4*) As shown in figure 1 this core was secured from the New Zealand plateau in 300 fathoms of water. There is no consistent trend in the radium content, in fact the radium content is most remarkably uniform. It is very doubtful that five samples spaced in like manner over 176 cms. in any non-marine formation would exhibit such a constant radium content.

One might assume that the rate of deposition on the shelf off New Zealand is extremely rapid. The expected increase in radium content might then be hidden in the errors of the measurements, but the age of the sediment at the bottom of core N-10 could hardly exceed 500 years if we employ the greatest spread in the results given in Table II. With due consideration for the distortion in the core sample this would imply a rate of deposition of at least 5 meters per thousand years.

It appears more likely that the sediments of the continental shelf consist largely of detrital material and that the operation of the mechanism that destroys the state of equilibrium between radium and ionium in deep-sea deposition is con-

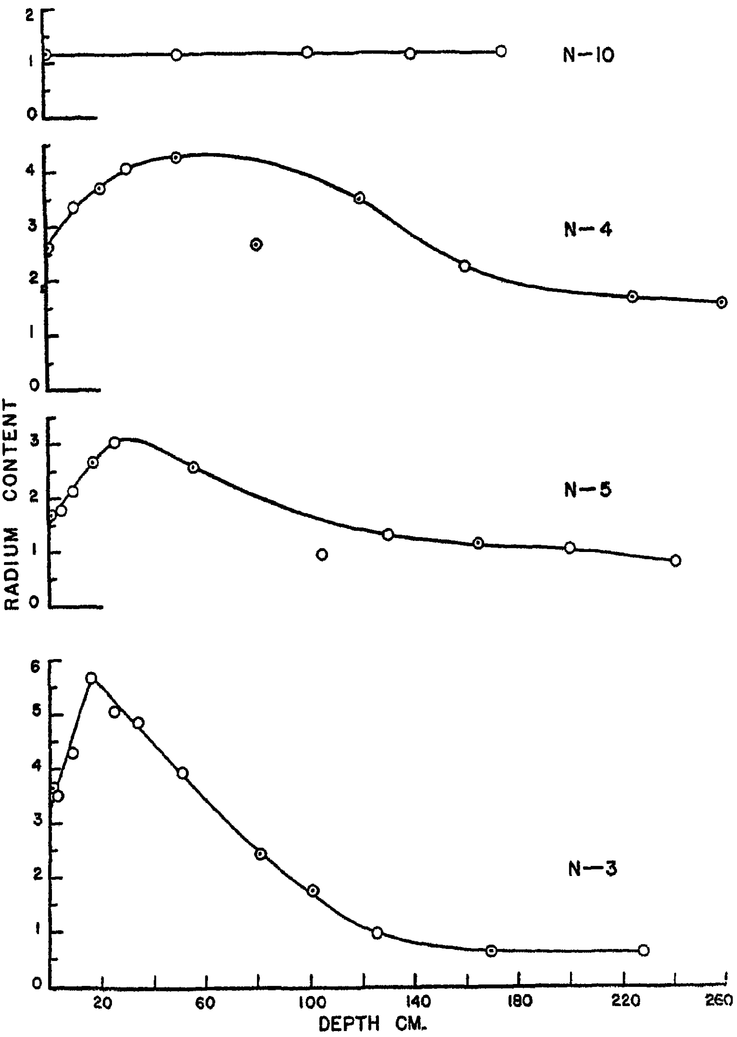

Figure 4 N-3, N-4, N-5 — The variation of radium content with depth in cores of glacial marine deposits of the Ross Sea, arranged in order of increasing rapidity of deposition reading upwards. N-10 — The variation of radium content with depth in a core of greenish gray mud from the New Zealand Plateau. The radium content is given in units of 10^{-12} g. radium per g of sediment dried at 110°C

fined to the deep-ocean basins.[16] If this is true, we are deprived of our method of measuring time on the continental shelf, but on the other hand we have a valuable clue to the nature of the mechanism by which the radioactive equilibrium is disturbed in that it requires deep water and the presence of only minor amounts of detrital material for its operation. From this, it follows that the deep-ocean sediments consist almost entirely of chemical and biochemical products. Such a conclusion seems obvious with respect to the calcium carbonate portion of a globigerina ooze, or to the silica of the diatoms or radialaria in these types of ooze; but there has been difference of opinion as to the formation of deep-sea clays devoid of organic skeletons, and as to the nature of the material that binds the organic skeletons together in those oozes consisting largely of organic remains.

ANALYSIS OF RESULTS

The purpose of this investigation was threefold.

Firstly, the results are being analyzed to determine the dates of important events in the Earth's recent history in the southern hemisphere. These events should be indicated by studies, now in progress, of the lithology and of the foraminifera of these cores. It is hoped that at least a tentative answer to the question of contemporaneous glaciation in the two hemispheres can be given by comparing this history with that derived from an earlier and similar co-operative endeavor on cores from the northern hemisphere.

Secondly, the rates of deposition of the sediments at the location of the cores from the southern hemisphere are in the course of evaluation both with respect to the type of sediment and to time in the past.

Thirdly, another analysis of the data is in progress which will add to the available information on the relations between the radioactive elements present in sea water and in ocean bottom sediments. A study of the initial concentrations of the radioelements in the surface of ocean sediments at the location of fourteen cores, including those described here, has revealed interesting correlations between the concentrations of the radioelements, the rate of their deposition, as distinct from the rate of deposition of the sediment itself, and the fineness of the sediment.

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[16] It is not possible to state that the mechanism causing a deficiency of uranium with respect to radium and ionium is also inoperative on the shelf. In order to demonstrate this it would be necessary to examine a very much greater depth of shelf-sediment than is represented by core N-10.

SUMMARY

Seventy determinations of the radium content of ocean sediments are reported here for a number of depths in each of six cores secured from the southern hemisphere. Curves of the radium content as a function of depth below the ocean bottom are similar in all qualitative respects to the curves previously obtained for cores from the North Atlantic and Caribbean Sea.

ACKNOWLEDGMENTS

It is a pleasure to acknowledge the cooperation of the Hydrographic Office of the U.S Navy in placing the necessary material at the author's disposal through the good offices of Dr. R. H. Fleming. Dr. J L. Hough contributed greatly to the success of the investigation, primarily in securing the core samples with the aid of his shipmates on the U.S.C.G.C. *Northwind*, and later by collaborating with the author in sampling the cores and assembling the necessary correlative data. The author herewith expresses his sincere thanks to Dr. Hough.

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DOLICHOHARPES AND THE ORIGIN OF THE HARPID FRINGE

HARRY B. WHITTINGTON

ABSTRACT A new genus of harpid trilobite *Dolichoharpes* (genotype *Eoharpes uniserialis* Raymond, 1925, from the Kimmswick Limestone of Missouri) is diagnosed and the type redescribed A second species, *D. reticulata*, from the Edinburg Limestone of Virginia, is also described. Specimens of the latter are silicified, and show the morphology of the cephalon in detail. The pits in the bilaminar fringe have flattened bases, but there is no evidence of an opening in the base, and the fringe was therefore apparently not perforate. The glabella and cheek lobes are ornamented by a prominent network of raised ridges surrounding depressions with flat bases, and this ornament merges into the upper lamella of the fringe.

In 1921, Richter suggested that the harpid fringe may have originated from funnel-shaped depressions originally present on the cephalic margin and the doublure. The depressions would be opposed to each other, and by deepening towards each other until the bases came into contact would develop into the characteristic pitted bilaminar fringe. The reticulate ornament of *Dolichoharpes*, if originally present on both the doublure and the dorsal surface, would form an ideal starting point for such a development. The structure of the cephalon strongly suggests that this was the case, and thus affords remarkable support for Richter's hypothesis.

INTRODUCTION

DURING the summer of 1947 the writer spent three months at the U. S. National Museum studying the collection of silicified trilobites from the Edinburg Limestone of Virginia. This visit was made possible by a grant from the Geological Society of America to Dr. G. Arthur Cooper. It is a pleasure to record my indebtedness to Doctor Cooper and to officials of the Society. Through the kindness of Dr. A. S. Romer the types of the trilobites from the Appalachians described by Dr. P. E. Raymond in 1925 were loaned to me for study and comparison with the silicified material. Among these specimens were the cotypes of *Eoharpes uniserialis* Raymond, 1925. My studies of European harpid trilobites showed me that Raymond's species was the representative of a new genus. Fragments of harpids occur very rarely in the silicified Edinburg material, and represent a different species of the same genus. From a block of the limestone, prepared in England, the incomplete cephalon, here described as the holotype of this new species, was obtained. Although this cephalon is incomplete and broken into four fragments, it is large compared with the majority of specimens extracted from this limestone.

The excellent preservation of the silicified cephalon reveals the details of the morphology. It is here suggested that these structures appear to confirm an hypothesis propounded in 1921 by Dr. R. Richter to explain the origin of the harpid fringe. In the following pages the new genus is diagnosed, the two species described, and the structure and origin of the fringe discussed. Reference is also made to the occurrence of a species of harpid in Northern Ireland, believed to belong to the new genus. This species, described long ago by J. E. Portlock, will be redescribed in a forthcoming monograph on British Harpidae.

Dr. G. Winston Sinclair is now studying the American Harpidae, and I am indebted to him for much helpful discussion and for his generosity in suggesting that I should publish this paper in advance of his work. My thanks are also due to the Research Committee, University of Birmingham, for grants in support of my work, and to Mr. L. W. Vaughan of the Geology Department for the preparation of the enlargements and prints from my negatives.

SYSTEMATIC DESCRIPTIONS

Family Harpidae Hawle and Corda, 1847

Genus *Dolichoharpes* Whittington, gen. nov.

Genotype· *Eoharpes uniserialis* Raymond, 1925

Diagnosis: Cephalon strongly convex and ovate in outline, the long axis being longitudinal. Glabella strongly convex, standing well above cheek lobes, highest point at mid-line of posterior margin, first[1] (basal) glabellar lobes large, triangular, first furrows deep, straight, second and third furrows very short and shallow. First glabellar lobes crossed by a prominent ridge running in a curve convex forwards. Between first and second furrows are less strongly ornamented oval areas demarcated by shallow furrows which curve round forwards and outwards from the inner ends of the first glabellar furrows. Eye tubercles and eye ridges well marked. Alae large, gently convex, depressed below level of adjacent parts of cheek lobes. Fringe with deep girder extending to tips of prolongation;[2] internal portion, notably cheek-roll prolongations, broad and

[1] Numbering lobes and furrows forward from the posterior margin of the glabella

[2] For terminology see Whittington "Redescription of *Eoharpes* Raymond, 1905," Geol Soc., London, Quart. Jour., 104, 221-8, pls. xi, xii, 1948.

vertical, external portion narrow with flat lower lamella and concave upper lamella. Marginal band broad. Pits coarse, irregularly and closely spaced in a reticulate pattern except in the upper lamella externally, where there is a tendency to an arrangement of the pits in radiating furrows separated by ridges. Coarse reticulate ornament on cheek lobes and glabella.

Thorax and pygidium unknown.

Stratigraphical Range and Distribution· Middle Ordovician of North America and Ireland. In this paper species of *Dolichoharpes* from the Kimmswick Limestone of Missouri and the Edinburg Limestone of Virginia are described. Material that I have examined in the U. S. National Museum suggests that the genus is also present in the Bromide of Oklahoma. *Dolichoharpes doranni* (Portlock), 1843, from the Bardahessiagh Beds of Pomeroy, Northern Ireland, seems also to belong to this genus. It occurs very rarely in beds which are most probably Middle Ordovician in age (cf. Stubblefield, 1939, p. 60).

Dolichoharpes uniserialis (Raymond), 1925

Plate 1, Figs 1, 3, 4, 5

[3]1925 Raymond, P. E., p. 15, Pl. 1, figs. 8, 9

Lectotype (here selected). MCZ (Museum of Comparative Zoölogy, Harvard College) 1681, an incomplete external mold of the cephalon.

Other material: MCZ 1682, the second of Raymond's original cotypes, an incomplete cephalon.

Description: The outline of the convex cephalon is oval, the greatest width being at about half the length. The maximum height (at the mid-line of the occipital ring) is about one-third of the width. The length of the prolongations is twice the axial length of the cephalon, and the tips of the prolongations curve inwards. The glabella tapers forwards to the rounded anterior end, the width being greater than the length. The length of the glabella is about one-quarter the length of the cephalon. The deep first glabellar furrows commence at about two-thirds the length and run straight inwards and backwards to isolate the large triangular first lobes. Shallow furrows connect the inner ends of the first glabellar furrows with the occipital furrow. From the adjacent median portion of the glabella a ridge runs across these connecting furrows and continues across the first glabellar lobes in a curving path, convex forwards (pl 1, fig. 1). The second glabellar furrows are shallow and short, situated at a point less than half the length of the glabella. From the inner end of the first glabellar furrows shallow furrows run forwards and outwards in a spiral path across the second lobes, finally curving round to run backwards and die out without reaching the first furrows at their outer ends. These furrows thus delimit oval areas which are part, but not all, of the second glabellar lobes. The third glabellar furrows are very short and shallow. The alae are semi-circular in outline, the length equal to about half the length of the glabella. They are gently convex, the outer margins sharply depressed below the adjacent cheek lobes, inwards rising to meet the shallow axial furrows. The occipital furrow is deep, curves forward to the mid-line, and is constricted behind the first lobes. The occipital ring is longest in the median line. The cheek lobes are gently convex in front of the alae, and steeply, almost vertically, bent down at the margins. The eye tubercles, poorly preserved in the type material, are situated opposite a point about one-third the length of the glabella and on the antero-lateral portions of the cheek lobes. The eye ridges run directly inwards to meet the axial furrows opposite the third glabellar furrows. The posterior marginal furrow is shallow, and merges into the alae. The posterior border is increasingly strongly raised laterally, becoming wall-like and continuous with the internal rim.

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[3] *D. uniserialis* was also described by Bradley (1930, Contr. Walker Mus., 2, No. 6, 244-246, pl 28, fig 8-10) but I have not seen this description or the material upon which it was based

The fringe is divided by the girder into the broad vertical internal portion and the narrower gently sloping external portion, the upper lamella of which is concave upwards, the external and internal rims are thick, and the broad marginal band has a median ridge, along which the suture presumably runs. MCZ 1682 (pl. 1, figs. 4, 5) has the upper lamella stripped off anteriorly and laterally, and shows that the inner margin of the lower lamella runs in a curved course around the postero-lateral margin of the cheek lobes and thence forwards and slightly upwards to pass in front of the glabella along the anterior margin of the pre-glabellar furrow. In the internal portion of the fringe the pits anteriorly and laterally are coarse and arranged in a reticulate pattern as far back as about one-third the length of the prolongations, while behind this point they become finer. In the concave depressed portion of the upper lamella the pits appear smaller and, especially laterally, are arranged in radiating rows of 2-3 pits, placed between broad radiating ridges. At the external margin pits appear on the radiating ridges and break up this arrangement. "Along the deepest part of the brim" there is not, as Raymond states a "single line of large pits."

The glabella and cheek lobes are ornamented with a coarse network of raised ridges. Raymond is not correct in stating that the glabella is smooth, and indeed he illustrates the ornament on his pl. 1, fig. 9. At the outer margins of the cheek lobes the ridges of the ornament merge into those between the pits of the upper lamella of the fringe, the boundary between the cheek lobes and the fringe being revealed by the greater relief in the ornament of the latter.

Geological Horizon and Locality: Kimmswick Limestone of Mincke, St. Louis County, Missouri.

Dolichoharpes reticulata Whittington, gen. et sp. nov.

Plate 1, figs. 2, 6, 7; Plate 2

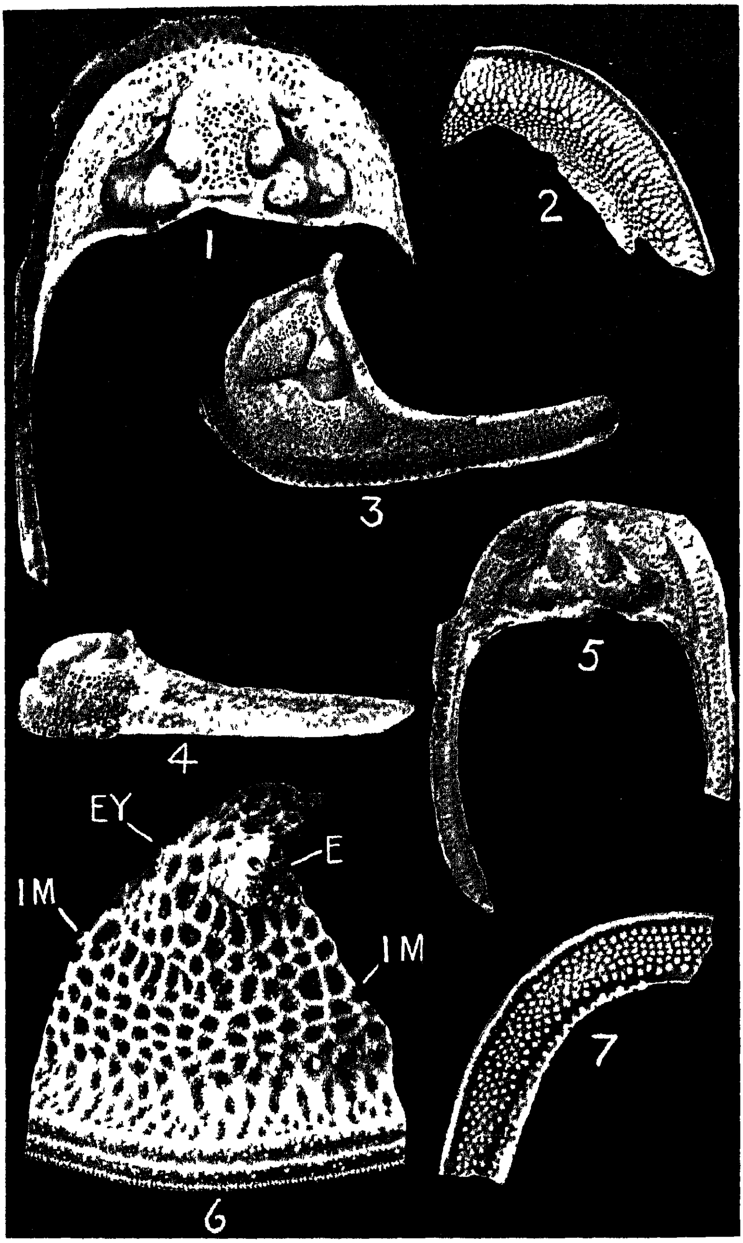
Holotype: The specimen illustrated in plate 1, fig. 6 and plate 2, in the U. S. National Museum.

EXPLANATION OF PLATE 1

Untouched photographs

Figs. 1, 3, 4, 5. *Dolichoharpes uniserialis* (Raymond). Kimmswick Limestone of Missouri. 1, 3. Dorsal and antero-lateral views of a rubber cast of the lectotype, MCZ 1681. Original of Raymond's Pl. I, fig. 9. x 3, x 2. 4, 5. Lateral and dorsal views of a second specimen, MCZ 1682, original of Raymond's pl. I, fig. 8. x 3.

Figs. 2, 6, 7. *Dolichoharpes reticulata* Whittington, gen. et sp. nov. Silicified specimens from Edinburg Limestone of Virginia. 2. Ventral view of left antero-lateral portion of cephalon, showing inner surface of upper lamella and adjacent portion of cheek lobe. x 3. 6. Left antero-lateral portion of holotype cephalon, to show the eye tubercle (E), the eye ridge (EY), and the ornament. IM is the position of the inner margin of the lower lamella, and thus indicates the boundary between the cheek lobe and the fringe. x 6. 7. Dorsal view of inner surface of left antero-lateral portion of lower lamella. The inner edge is broken and the inner margin is not shown. x 4½.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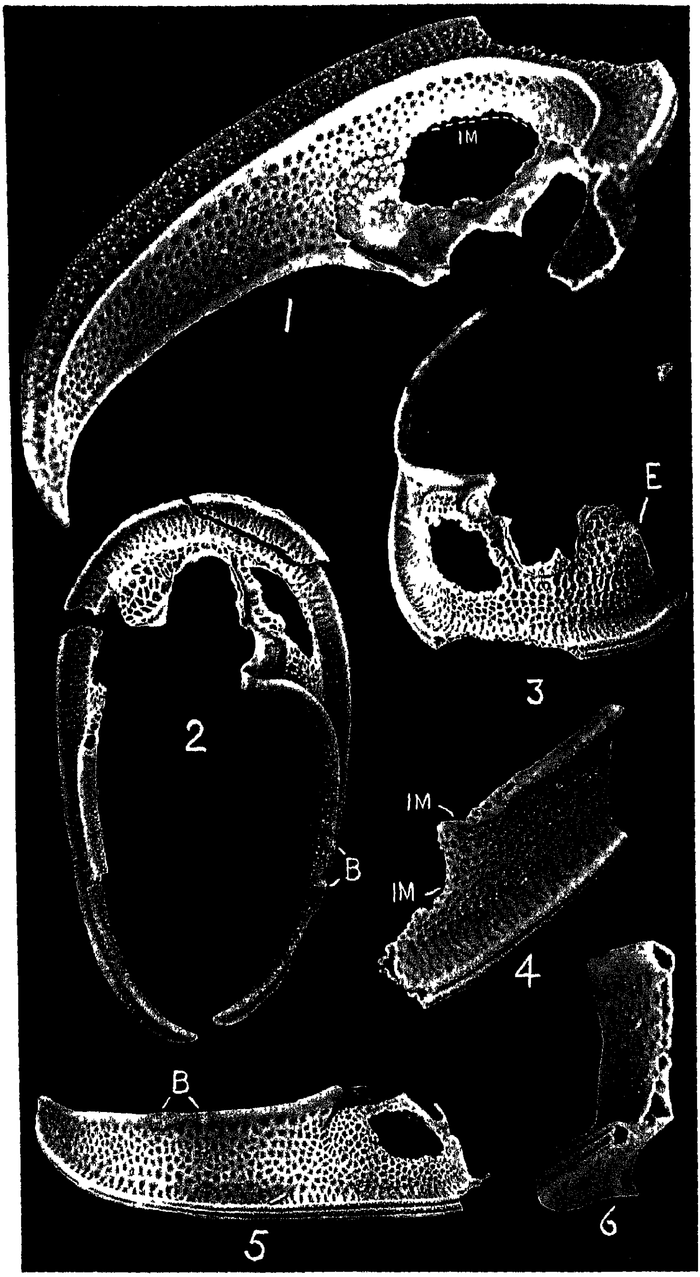
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Other material: two incomplete cephala and various fragments of the fringe, also in the U. S. National Museum.

Description: The glabella of *Dolichoharpes reticulata* is not preserved in the holotype or in any other specimen so far obtained. This species, however, resembles *D. uniserialis* in all other features of the cephalon—outline and shape, the form of the alae, the structure of the fringe, the ornament of the cheek lobes, the position of the eye tubercles—and it is therefore placed in the same genus.

Dolichoharpes reticulata may be distinguished from *D. uniserialis* by differences in detail between the fringes of the two species. In *D. uniserialis* the internal portion of the fringe is inflated anteriorly, and descends vertically to the girder (pl. 1, figs. 1, 4). In *D. reticulata* the corresponding portion of the fringe is not inflated, and slopes steeply down to the girder but is not vertical (pl. 2, fig. 5). The internal rim in *D. reticulata* is very prominent and extends horizontally back from the posterior margin (pl. 2, fig. 5). In *D. uniserialis* the internal rim appears less prominent and slopes gently down and back from the posterior margin (pl. 1, fig. 4). Thus in *D. reticulata* the internal portion of the fringe maintains a greater width posteriorly than in *D. uniserialis*. The external portion of the fringe in *D. reticulata* is wider anteriorly and laterally than in *D. uniserialis*, the pits are more numerous and, while radiating ridges are developed on the upper lamella of *D. reticulata*, they are not so broad as in *D. uniserialis*.

EXPLANATION OF PLATE 2

Untouched photographs

Figs. 1-6. *Dolichoharpes reticulata* Whittington, gen. et sp. nov. Holotype, a silicified specimen broken into four fragments, from the Edinburg Limestone of Virginia. 1. Ventral view of largest fragment, showing lower lamella of fringe with girder The line IM indicates the inner margin of the lower lamella. Within this line may be seen the ventral surface of the right cheek lobe and ala x 3. 2. Dorsal view of holotype, the four fragments arranged in approximate relative positions. The larger fragment is shown in figs. 1, 3, 5. B indicates a bryozoan growing on the internal rim. x 2. 3, 5. Anterior and lateral views of the largest fragment of the holotype E on fig 3 is the eye tubercle shown on pl I, fig. 6 B is a bryozoan growing on the internal rim. x 2. 4, 6. Antero-lateral and posterior views of the larger (anterior) fragment of the left prolongation of the holotype, to show the bilammar fringe in section IM indicates the position of the internal margin of the lower lamella, within which is a small portion of the postero-lateral corner of the left cheek lobe. x 3, x 4½.

The beautifully preserved silicified specimens of *Dolichoharpes reticulata* reveal the structure of the fringe, the ornament, and other details. The eye tubercle (pl. 1, fig. 6 E) is seen projecting upwards and outwards over the steeply-sloping antero-lateral portion of the cheek lobe. The two raised oval areas (the anterior has adventitious material clinging to it, the posterior is partially broken) are identical in appearance with the two lenses shown to exist in *Harpes macrocephalus* (cf. Richter 1921, p. 194-5, pl. 16, fig. 4, pl. 17, fig. 10). A central circular raised area is also present, corresponding with that in *H. macrocephalus* which is said by Richter not to be a lens. The reticulate ornament on the cheek lobes consists of high ridges enclosing flattened areas. Upon the ridges are scattered small tubercles, those at the intersections being larger and more prominent. The reticulate ornament merges into the pitted fringe structure and the boundary between the cheek lobe and the upper lamella is not easily discerned (pl. 1, fig. 6). The ventral view (pl. 2, fig. 1) reveals the inner edge of the lower lamella curving from the internal rim around the postero-lateral corners of the cheek lobes and running anteriorly just outside the pre-glabellar furrow. The girder is deepest anteriorly, and the rims are prominent, the marginal band having the median ridge along which the suture runs. The sections (pl. 2, figs. 4, 6) show how the flat external portion of the lower lamella is combined with the concave upper lamella. The pits narrow inwards and have a flat base, the bases of opposing pits being in juxtaposition, and the suture passing between them.

In complete specimens of the fringe, or in separate, single lamellae, the base of the pits is represented by a thin layer of silica, much thinner than the walls of the pits or the areas between the pits. Only rarely is seen a tiny hole connecting opposing pits, and where this occurs it is apparently due to breakage or abrasion of the specimen. Since the material is so perfectly silicified, and very fine details of the ornament are preserved, it seems reasonable to infer that in the living animal the bases of the pits were covered by a thin layer of chitin. The structure of the fringe is therefore exactly as described by Richter (1921, pp. 184-7, pl. 16, figs. 4, 5) except that there do not seem to be openings connecting opposing pits. Richter's material was of Devonian age, and it seems that either the structure of the pits in Devonian harpids was different from that of Ordovician ones, or that the thin layer at the base of the pits was not preserved in Richter's specimens.

An ornament of fine tubercles occurs on the inner and outer rims, and the marginal band. Tubercles occur at the junction of the ridges separating the pits on the upper lamella of the cheek roll and the external portion of the lower lamella.

Geological Horizon and Locality: Basal Edinburg Limestone (? Botetourt member), Middle Ordovician. Holotype from section in field on south side of road and 0.2 miles east of Strasburg Junction, Shenandoah County, Virginia, from crags of dark granular and fine grained limestone about 155 feet southeast of the edge of the quarry dump at the top of the field. Other material, including the two fragments shown in pl. 1. figs. 2, 7, from bed 18 in Geologic Section 19 of Cooper and Cooper (1946, p. 94-5).

STRUCTURE AND ORIGIN OF THE HARPID FRINGE

The structure of the fringe, as described above, appears identical with that of Devonian harpids described by Richter (1921) except that there is no opening, but a thin layer of silica at the base of the pits in the silicified specimens. Extremely fine detail is preserved in tiny specimens from these limestones, and the chitin-replacing silica is very fine-grained. I regard it as unlikely that the silica has grown as a thin layer across what in the living animal was a tiny opening not covered by chitin. Only fragments of young cephala have so far been found but in these the silica layer is of a constant thinness over the glabella, cheek lobes and fringe including the pits, with no suggestion of any opening at the base of the pits.

As has long been recognized, the structure of the harpid fringe is remarkably similar to that of the Trinucleidae (see e.g. Richter, 1921; Størmer, 1930; Whittington, 1941 a, b). The following points of difference may be noted: (a) the canals seen in the marginal part of the cephalon of *Tretaspis* and *Trinucleus* (Størmer, 1930, pp. 100-102) have not been observed in the Harpidae; (b) terrace lines have not been observed on the fringe of harpids; (c) in the Trinucleidae it is believed that the fringe is perforated at the base of the pits (Størmer, 1930, p. 102-103; Whittington, 1941b, p. 509).

The evidence of the silicified Trenton *Cryptolithus* is, however, equivocal, partly because the silicification is coarser than that of the Edinburg material. It seems then, that the Harpidae may include genera with perforate and genera with imperforate fringe, and the same may be true of the Trinucleidae. This condition is theoretically possible if the mode of origin of the fringe discussed below be accepted. It would seem, however, that the adaptive value of a perforate and a nonperforate fringe would be different.

The origin of the fringe and the nature of similar structures in crustaceans and in the tortoise beetle (*Cassida*) have been discussed by Richter (1921, p. 188-191) One of the possible modes of origin of the fringe suggested is as follows (p. 189): if funnel-shaped depressions existed in the borders of the cephalon and in the doublure, opposing pairs of funnels might deepen towards each other, the bases eventually coming into juxtaposition, and finally the funnels opening into each other to form an hour-glass shaped connection between the opposing lamellae. The characteristic ornament of the glabella and cheek lobes of harpids is the raised reticulate pattern of ridges surrounding depressions with flattened bases. Such ornament, extending onto the cephalic borders and the doublure, would form an ideal starting-point for the evolutionary trend suggested by Richter. I believe that *Dolichoharpes* affords clear evidence that this is what actually happened in the Harpidae. The reticulate ornament is strongly developed, and as has already been pointed out, it is difficult, on the dorsal surface, to perceive the boundary between the ornamented cheek lobes and the upper lamella of the fringe, and the similarity between the ornament and the pits in the upper lamella is obvious (see pl. 1, fig. 6; pl. 2, figs. 2-5). It seems that in *Dolichoharpes* the bases of the original depressions in the two lamellae forming the fringe have come into contact and been flattened against each other but do not open into each other. Apparently in the geologically younger species *Harpes macrocephalus* this latter stage is reached, and it may be noted that in this species the ornament of the cheek lobes and the glabella is of small scattered tubercles (Richter, 1921, pl. 16, fig. 3a; pl. 17, fig. 10). The purpose of the pitted structure is clearly to stiffen the thin, bilaminar fringe, as Richter has said (1921, p. 209-210). The development of the thickened rims, the girder, and radial ridges on the upper lamella also promoted this rigidity.

Reference has been made to the similarity in structure of the harpid and trinucleid fringe. Reticulate ornament is equally characteristic of the cheek lobes and glabella of the Trinucleidae (see Størmer, 1930; Whittington, 1941a), and it is possible that the pitted fringe originated in the same way as suggested for the Harpidae. It seems to me that this similarity of structure and origin need not necessarily be regarded as implying that the two families are closely related.

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REVIEWS

The Essentials of Organic Chemistry; by C. D. PORTER and T. D. STEWART. Pp. vi, 394. Boston, 1948 (Ginn and Co., $4.00).—This book is designed to fill the need for a text to be used by pre-medical, pre-dental and pre-nursing students, as well as by students in the liberal arts, in a semester course in introductory organic chemistry. The text, written by the same authors as the earlier and more complete *Organic Chemistry* (Ginn and Co., 1943), presents a survey of the essential chemistry of the aliphatic and aromatic compounds, and special brief chapters are devoted to the carbohydrates, the proteins and amino acids, dyes and indicators, natural and synthetic drugs, the enzymes, vitamins and hormones, and to the sterols. Stereochemistry is introduced at an appropriate location, and theoretical material, such as the nature of the chemical bond and resonance, is reduced to its barest descriptive essentials. The relations and applications of organic chemistry to the biological sciences and to industry are concisely and lucidly illustrated. Each chapter is accompanied by a set of exercises and a set of problems which have been selected with due consideration of the mastery which a student can be expected to attain in such a courrse. It is the opinion of this reviewer that the authors have achieved their task of condensation and simplification with a minimum sacrifice of useful knowledge, and that the text can be used with every expectation of success. ALFRED H. FRYE

Manual of Coastal Delineation from Aerial Photographs; by P. G. MCCURDY. U. S. Navy Department, Hydrographic Office Publication No. 592, Pp. iii, 143, 202 figs. Washington, D. C., 1947 (The Hydrographic Office, Washington 25, D. C., $1.50).—This manual is intended primarily, as the author states, to make photogrammetrists, hydrographic engineers, and cartographers conscious of each other's problems, to help those delineating coastal areas from aerial photographs, and to help those using the delineated photographs in compiling coastal charts The manual, nevertheless, has definite geological appeal, and is very useful in geologic instruction It is well organized, the diagrams and photographs appear on pages opposite the discussions to which they pertain, and reproduction of the photographs is excellent. The text is typed with right-hand margin control, and both text and photographs are offset reproduced. Of the 202 figures, 162 are aerial photographs (97 vertical stereopairs and 59 obliques); 25 are diagrams and 15 are chartlets (small nautical charts with soundings shown in fathoms). Many of the instructive diagrams are taken from Johnson's *Shore Processes and Shoreline Development*, 1938, and Lobeck's *Geomorphology*, 1939.

The four major coastal types, explained in detail with accompanying diagrams, chartlets, and oblique photographs, are those proposed by Johnson in 1938. Two other types of coasts considered and pictured are the vegetation-bordered coast (capably described in *Pacific Landforms and Vegetation*, U. S. Navy Photographic Intelligence Center Report No. 7, 1945), and the man made coast. A seventh type, the ice barrier coast, is to be described in a later edition. From the viewpoint of persons concerned with the description of a coast and not too greatly interested in its geologic history, it would seem logical and simpler to use a descriptive classification of coasts, instead of the genetic approach of Johnson, which requires field evidence for final proof. Because of this requirement, some of the coasts described in the manual, based only upon photographic interpretation, are subject to correction. A descriptive classification also would be in keeping with the well-founded vegetaton-bordered coast, the man made coast, and the ice barrier coast.

The chapter "Coastal Items To Be Mapped" is beautifully illustrated with vertical stereopairs. It is actually a chapter in beach interpretation from aerial photographs, and is suitable for teaching geomorphology and photographic interpretation. The chapters on "Manuscript Symbols" and "Recommended Procedure for Delineation of Coastal Features" are for the cartographer and photogrammetrist. The final chapter illustrates the problems in delineation of coastal features by carefully selected and arranged oblique photographs, with duplicate sets of vertical stereopairs of part of the same area shown on the page opposite The first set of stereopairs is untouched, the duplicate set is marked to show the delineation of the particular coastal features under discussion.

From a geological standpoint it is regrettable that several small errors of usage and misapplication of terms exist After careful scrutiny, the reader will find that the operational mechanics of longshore currents are not fully understood by the author As many photographs are of great geologic interest, it is disappointing that their localities are not given. The manual is valuable for teaching the physiography of coasts and shoreline features to elimentary classes in geology and geomorphology, and it serves as a worthwhile supplement to courses in aerial photographic interpretation and photogeology.

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American Journal of Science

MAY 1949

STUDIES OF SUPERGLACIAL DEBRIS ON VALLEY GLACIERS

ROBERT P. SHARP

ABSTRACT. On the stagnant ice of Wolf Creek Glacier, in the Canadian Yukon, fully 90 per cent of the superglacial debris is derived from an englacial source. A mantle averaging two feet thick has developed by ablation of ice containing but a fraction of 1 per cent debris. This mantle is a heterogeneous mixture ranging from large angular blocks 15 to 20 feet in diameter to well-bedded pond deposits of clay and silt. With increasing time in the superglacial environment size and angularity of fragments are reduced, sorting is improved, and mean grain size is increased through removal of fine material by melt water. On areas of ice long stagnant, 50 to 60 per cent of the superglacial debris has been reworked by running water.

A heterogeneous and irregularly distributed superglacial mantle promotes differential ablation which is a major factor in producing and maintaining the irregular topography of stagnant ice. Layers of fine debris, even though moist, provide insulation equivalent to a foot or two of coarse debris owing to the ease with which air and water circulate through the larger openings of the latter. Ice well insulated by superglacial detritus forms irregular mounds and ridges which increase in size until progressive thinning of the debris mantle, or its removal by some independent agent, permits more rapid ablation, eventually resulting in a complete inversion of topographic relief. Under conditions now existing in the Wolf Creek area, decades are probably required for such an inversion in the areas of heavy superglacial mantle. Glacial tables, moraines, mud ridges, and other features of the superglacial mantle are also described briefly.

Dust wells and dust basins form on relatively clean ice where indirect ablation exceeds direct ablation. In this region much of the bottom silt is probably not dust. Continued deepening of wells with shaded bottoms is attributed chiefly to direct absorption of diffused radiation and partly to transmission of direct and diffused radiation through the ice. Dust basins are more largely the product of direct radiation, and their consistent east-west elongation is ascribed to greater combined retreat of the east and west walls than of the north wall alone. The division of these features into alpine- and polar-type *Kryokonitlocher* does not hold on Wolf Creek.

The complete wasting of stagnant ice leaves an accumulation of debris reproducing on reduced scale the irregular topography of the stagnant ice and having the constitution of the superglacial mantle.

INTRODUCTION

A SIGNIFICANT part of the detritus left by many glaciers is of superglacial origin, and its nature is partly a product of that environment. Superglacial debris also influences glacial behavior, particularly in the stage of deterioration (Ogilvie, 1904, pp. 723, 725, 742; Flint, 1942, pp. 117-119), and it is a major factor in the formation of a variety of features on the ice surface. For these reasons, a more complete treatment of data previously published in abbreviated form (Sharp, 1947) seems desirable.

Field studies were made between June 26 and September 6, 1941, as part of the scientific program of the fourth Wood Yukon Expedition sponsored by the American Geographical Society. The assistance of all members of that expedition and especially of its leader, Walter A. Wood, is warmly appreciated. Research grants from the graduate schools of the University of Illinois and the University of Minnesota are acknowledged. The critical comments of Richard H. Jahns and the draftsmanship of Joan T. Rounds contributed to this study.

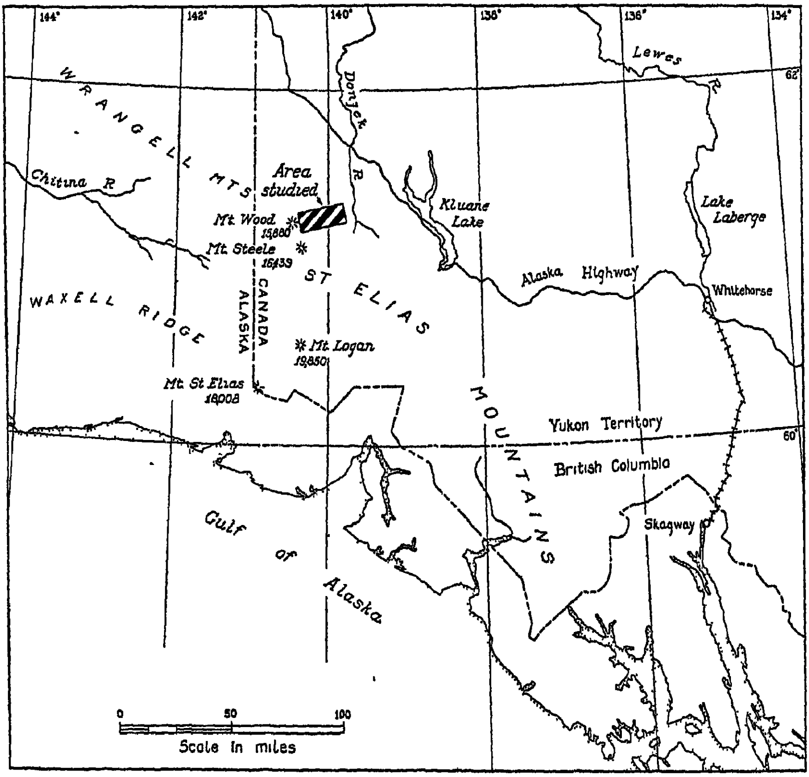

Fig. 1. Map of Wolf Creek area and environs.

PHYSICAL SETTING

The Wood expedition was conducted in the Wolf Creek region on the northeast flank of the St. Elias Range. This is in the Yukon Territory of Canada about 25 miles east of the Alaskan border at 61° 15′ N. Lat. and 140° 10′ W. Long. (fig. 1). In this region the St. Elias Range trends northwestward along the Alaska-Yukon boundary and is a rugged, glacier-bearing mountain mass with numerous lofty peaks culminating in Mt. Logan (19,850). Wolf Creek Glacier extends 25 miles eastward from 16,000-foot peaks at its head to a terminal altitude of 4000 feet. The lower 9 miles of this glacier are stagnant, but the upper part and most of the tributary ice streams are active; some receding and some advancing in 1941. This provided opportunity for observing debris on ice in different conditions of activity.

Wolf Creek is situated on the interior side of a high mountain range, so its climatic environment is distinctly continental, in fact almost semi-arid at the lower elevations. The climatological data in table I show a mean annual precipitation of 9 to 11 inches, a maximum temperature in the high seventies or low eighties, Fahrenheit, and a mean annual temperature between 30 and 40 degrees for the Yukon Plateau east of Wolf Creek.

TABLE I

Climatological Data for Stations on the Yukon Plateau.

Station	Location with respect to Wolf Creek	Length of Record (years)	Mean Annual precipitation (inches)	Maximum Temperature recorded (°F)	Mean Annual Temperature (°F)
Burwash Landing	40 miles east	5	9.91	79	—
Whitehorse	175 miles east	9 (broken)	11.40	84	38
Carcross	200 miles east-southeast	31 (broken)	9.10	89	29
Atlin (Reed, 1941 p. 672)	250 miles southeast	18	10.94	86	—

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[1] Data on all stations except Atlin kindly furnished by the Weather Central of the 7th Weather Group, U. S. Army Air Forces

Much of the Wolf Creek area is probably not so dry, and the temperatures are certainly somewhat lower than at the above stations. Nevertheless, the greater height of orographic snowline here, 8800 to 9100 feet, compared to its altitude on the moister coastal side of the St. Elias Range, 2000 to 3000 feet, is indicative of the marked difference in climatic environment on opposite sides of the range. Wolf Creek Glacier suffers further from the fact that it flows eastward into areas of increasing aridity where clear skies and long hours of sunshine in summer produce extensive ablation. The climatic environment of this area must be kept in mind when making comparisons with features of ablation in other regions.

SUPERGLACIAL DEBRIS

SOURCES

Possible sources of superglacial debris are (1) material from extra-glacial areas dumped directly onto the ice below the firn line, and (2) englacial material brought to a superglacial position through lowering of the ice surface by melting. This englacial material, in turn, may be (a) extra-glacial debris dumped onto the ice above the firn line, (b) subglacial material brought to an englacial position by movement along shear planes, or (c) the ground moraine of inset and superimposed ice streams (Visser, 1932, pp. 34-39; 1935, pp. 23-24; Klebelsberg, 1938, pp. 25-39; 1939, pp. 310-311).

Most debris on Wolf Creek glaciers consists of material, formerly englacial, that has been concentrated on the surface by ablation. This is demonstrated by ice streams near the center of the trunk glacier, far from possible extra-glacial sources of debris, at least below the firn line, which have clean surfaces in their upper reaches and heavy debris mantles farther down-valley. Ten per cent or less of the detritus on Wolf Creek glaciers has been dumped onto the ice below the firn line from extra-glacial sources. By contrast, the mantle on Tasman Glacier in New Zealand (Speight, 1940, p. 134) is said to consist largely of material that has always been superglacial.

Concentrations of large angular blocks (pl. 1, fig. 1) were probably carried directly onto the glacier's surface by a rock fall or avalanche; but this may have occurred above the firn line, and the blocks could have been subsequently incorporated within the body of the glacier. Therefore, large size and angularity do not necessarily indicate solely a superglacial history. Streams from tributary valleys have built small alluvial cones onto the surface of Wolf Creek Glacier near its terminus. These cones are exceptional, purely local concentrations of superglacial material formed only where the glacier surface has been lowered by melting so that tributary streams can flow directly onto it.

Melting of at least 500 feet of ice in the lower few miles of Wolf Creek Glacier has produced a superglacial debris mantle that would be not more than 2 feet thick if uniformly distributed. Some detritus has been carried from the glacier by meltwater and some has fallen into crevasses and other englacial openings, to be sure, but even if a one-third loss by such means is assumed, only a fraction of one per cent of the melted glacier consisted of debris. The ice exposed at present near the glacier's snout contains much more detritus (pl. 3, fig. 1), but it must be remembered that the bottom, debris-rich part of the glacier is seen here.

NATURE OF SUPERGLACIAL DEBRIS

Debris on stagnant ice is accorded major consideration here as it is abundant in this area and is the type most likely to be a recognizable constituent of glacial deposits. One must keep in mind that detritus on active ice may be considerably different. Many previous descriptions of superglacial material (Garwood and Gregory, 1898, p. 208; Ogilvie, 1904, p. 732; Tarr and Martin, 1914, p. 61; Ray, 1935, pp. 314, 315) emphasize the coarse, angular, ill-sorted nature of the rock fragments. Although this may be true for debris on active ice, it is not wholly characteristic of the material on stagnant ice in this area. The mantle on stagnant ice contains many rounded stones (pl. 1, fig. 2), considerable well-sorted debris, and some well-bedded deposits (pl. 2, fig. 2). Even so, heterogeneity is the rule, for the debris mantle on ice long stagnant consists of material that has had an extended superglacial history and shows considerable modification by superglacial processes, of material that has but recently been brought to the superglacial environment by melting and is little modified, and of material in all intermediate stages. Erratic distribution and localized accumulations of different materials also contribute to this heterogeneity, for small deposits of well-rounded, well-sorted debris are isolated in the midst of larger accumulations of angular, unsorted fragments. Furthermore, the various ice streams composing a valley glacier not uncommonly carry debris differing lithologically from that of neighboring ice streams, and these differences are apparent in the superglacial mantle.

The debris mantle on Wolf Creek glaciers ranges from a sparse sprinkling on active ice to a nearly continuous cover 1 to 2 feet thick on stagnant ice. Somewhat greater thicknesses have developed near glacier snouts, on lateral margins, and along debris-rich bands in the ice, but such differences are overshadowed by local variations related to the topography of the ice surface (pl. 3, fig. 2). If the mantle were uniformly distributed, its average thickness would not exceed two feet, a figure notably less than the 5 to 10 feet cited by Tarr and Martin (1914, p. 206) or the 15 to 20 feet noted by Tarr (1908, p. 86) on coastal Alaskan glaciers.

The fragments in this debris range from clay and silt to huge blocks 15 to 20 feet in diameter. A casual observation gives the impression that the material is coarsest in the upper reaches of a glacier and finest near the snout. This is obviously true for the larger stones, but the progressive loss of fine material carried off by meltwater is so great that the mean diameter of the remaining debris probably increases. This conclusion is based solely on field observations and not on statistical laboratory analysis.

The size of superglacial fragments is reduced by weathering, particularly frost action, by impact of falling and rolling stones (Russell, 1893, pp. 226-227; Ray, 1935, pp. 315-316), and by the action of running water. The efficacy of weathering is demonstrated by abundant rock chips around boulders on the ice and by the shattering of some bedded and foliated rocks. Falling and rolling stones can be seen and heard in the areas of rough topography on stagnant ice (pl. 3, fig. 2), especially on warm days, and the effects of this process are shown by percussion marks and chipped areas on some boulders, and by the shattering of others. Streams of running water also are common, and the effects of abrasion during water transport are clearly evident.

Reduction in size is accompanied by a decrease in angularity, and even though much debris on stagnant ice is noticeably angular, considerable blunting of edges and many well-rounded stones can be seen (pl. 1, fig. 2). Opposing forces are at work in this respect, for frost action and breaking by impact increase angularity, whereas the actions of most transporting agents, particularly running water, reduce it. On Wolf Creek, at least, rounding by water transportation has the upper hand.

Degree of sorting increases with time in the superglacial environment. In summer, the superglacial material is shifted by running water, by mass movement, and by free fall initiated and controlled by differential melting and stream cutting. In addition to the sorting obviously produced by superglacial water, crude sorting is accomplished in other ways. A crevasse, narrowing downward, sorts material that falls into it by letting the smaller sizes fall to deeper levels. This sorting in a vertical plane can be transferred to the surface as the glacier melts and different levels of the debris-choked crevasse are exposed. Linear accumulations of large boulders on Wolf Creek Glacier are ascribed to this process. Crude sorting at the foot of debris-mantled ice slopes, observed here and elsewhere (Russell, 1893, pp. 226-227; Ray, 1935, pp. 315-316), is attributed to the fact that large stones roll farthest and most frequently. The first is a matter of momentum, and the second of instability related to perching on ice pedestals. Small stones become firmly seated by melting into the ice. The above mechanism also accounts for the observed concentration of large stones in low places.

The work of running water in dissecting the surface of stagnant ice and in transporting and reworking large parts of the superglacial mantle is considerable. The abundance of water-worn and water-stored superglacial debris shows that running water is an agent of the superglacial environment that must not be underestimated. Not only are many streams actually observed, but long sinuous deposits of rounded stones and sorted sand or gravel mark the abandoned courses of older streams. Many of these channel deposits have now been converted to ridge cappings by differential melting. The longer debris remains in the superglacial environment, the greater the opportunity for reworking by running water, and 50 to 60 per cent of the debris on lowermost Wolf Creek Glacier shows evidence of modification by water.

The surface of many masses of stagnant ice is dotted with ice-walled ponds (pl. 3, fig. 1), in which are deposited layers of well bedded sand, silt, and clay containing occasional stones and interbeds of coarse angular gravel (pl. 2, fig. 2). The coarse debris is supplied by free fall and mass movement from the mantle on surrounding ice slopes, and finer material comes from this source and from streamlets draining into the pond. In the instance of small ice-walled ponds, there is no need to call upon ice rafting as a source of coarse debris. The impact of large stones falling from the surrounding walls onto soft pond sediments produces small folds (pl. 2, fig. 2) with the intensity of deformation decreasing outward from the center of impact. Contemporaneity of these structures is demonstrated by truncation prior to deposition of overlying beds.

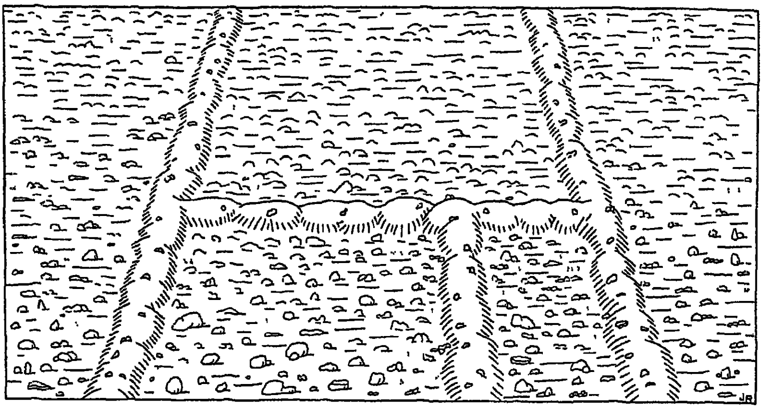

Fig. 2 Field sketch of mud ridges in superficial mantle of a Wolf Creek glacier. Ridges are 12 to 18 inches high, and area viewed is about 15 feet wide.

Striated stones are said to be rare in much superglacial debris (Garwood and Gregory, 1898, p. 208; Russell, 1893, p. 203; Ogilvie, 1904, p. 737; Ray, 1935, p. 317), but this is not true of detritus on Wolf Creek glaciers and on some Greenland glaciers (Salisbury, 1896, p. 806). The probable sources of such stones are: (1) Subglacial debris sheared up into the ice. (2) Ground moraine of inset and superimposed ice streams. (3) Purely englacial material scratched by shearing along planes within the ice (Garwood and Gregory, 1898, p. 223; Philipp, 1920, pp. 526-527). Most striated

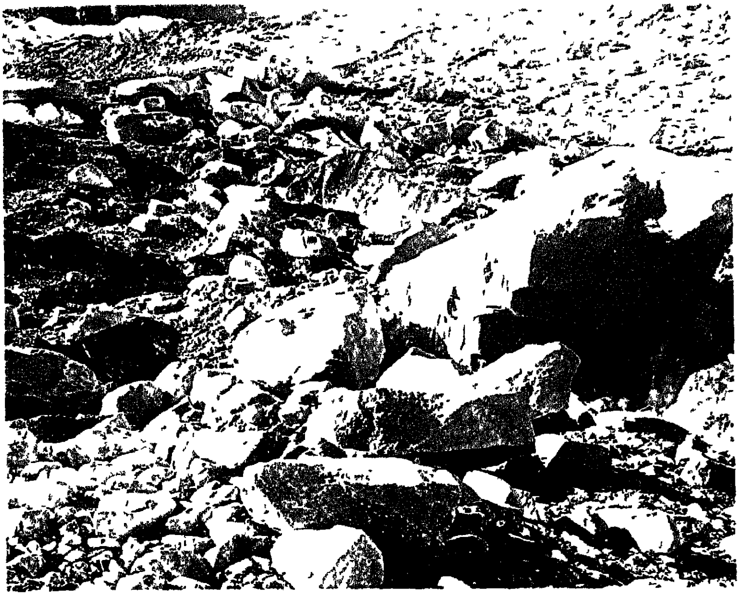

Fig. 1. The largest of these blocks on the surface of upper Wolf Creek Glacier is about 15 feet across, and all are composed of the same type of granite.

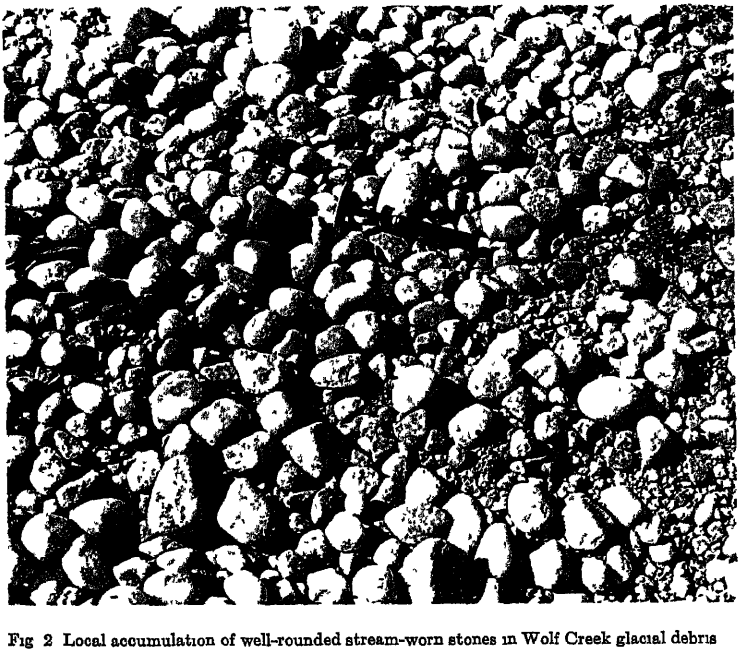

Fig. 2. Local accumulation of well-rounded stream-worn stones in Wolf Creek glacial debris.

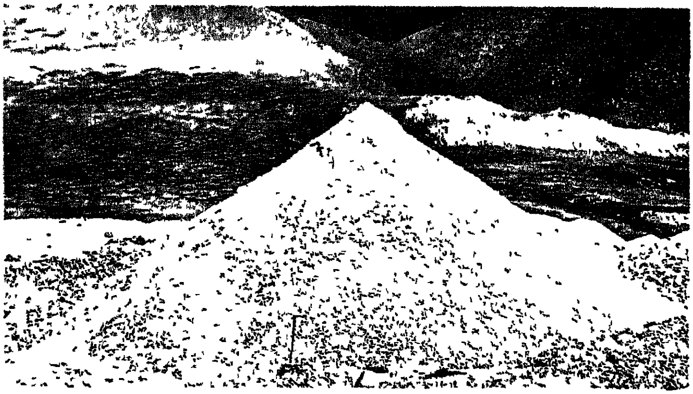

Fig 1. Debris-covered ice cone formed because of insulation provided by a localized deposit of relatively well-sorted gravel

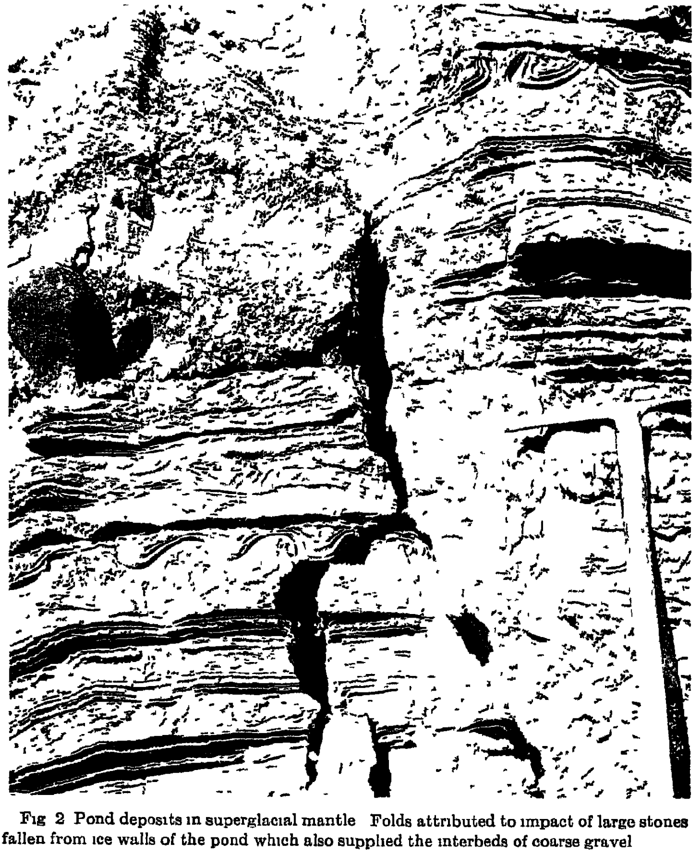

Fig 2 Pond deposits in superglacial mantle Folds attributed to impact of large stones fallen from ice walls of the pond which also supplied the interbeds of coarse gravel

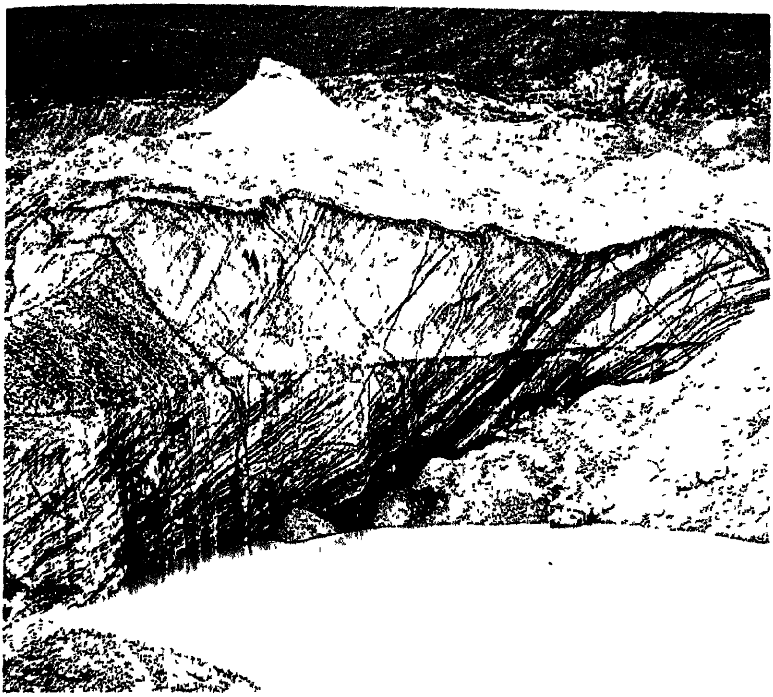

Fig 1 Pond on surface of lower Wolf Creek Glacier Also shown are abandoned shorelines, dipping debris bands in the ice, a debris-mantled ice mound in the background, and the thinness of the extensive superglacial mantle Shoreline on ice face is undercut

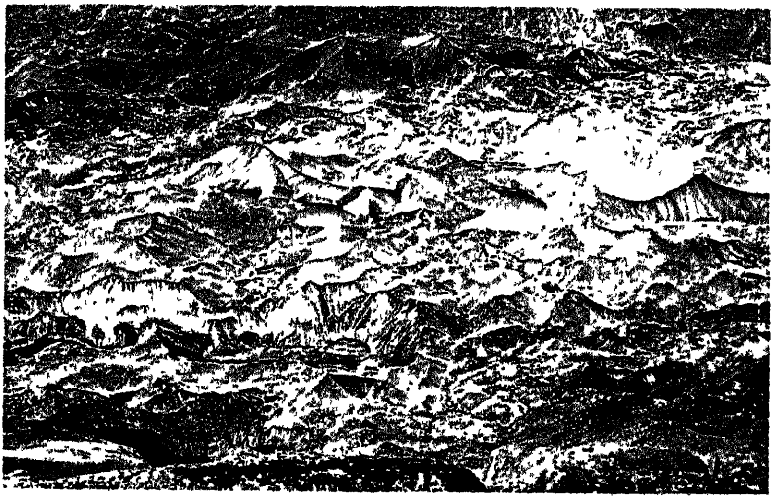

Fig 2 Chaotic topography and superglacial mantle of lower Wolf Creek Glacier viewed from several hundred feet above on the north bank.

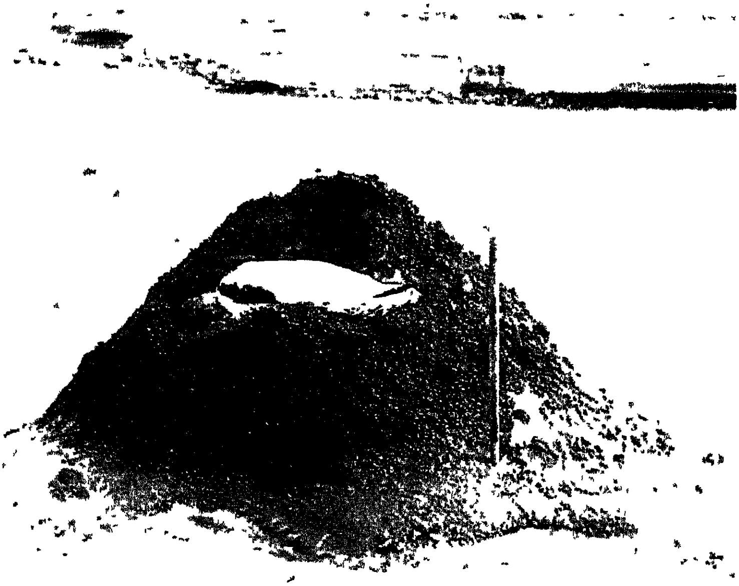

Fig 1 Debris-covered ice cone on annual river ice of Wolf Creek area pierced by small tunnel developed by melting where mantle was removed

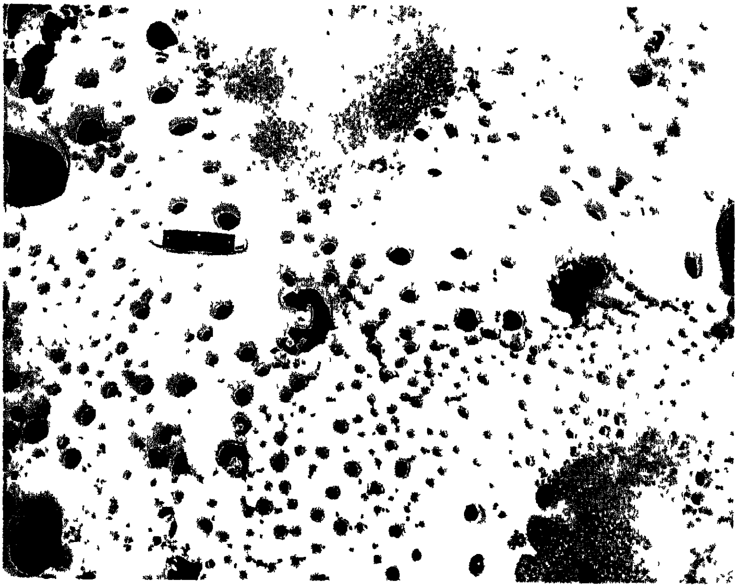

Fig 2 Small dust wells a few inches deep perforating surface of upper Wolf Creek Glacier

stones on Wolf Creek Glacier probably come from the ground moraine of inset ice streams which are believed to be an important element in the trunk glacier (Sharp, 1947, pp. 28-30).

A few low rounded ridges of mud have been observed in the midst of coarse, ill-sorted debris on relatively level areas of stagnant ice. The ridges are 12 to 18 inches high, straight to gently curving, 5 to 20 feet long, and form intersecting sets in plan (fig. 2). The mud is rich in clay and silt and contains angular stones which are concentrated as a thin residual mantle on the surface. Excavations reveal steep parallel walls at depth, and lateral spreading near the surface (fig. 3). In favorable exposures the ridges are seen to mark the traces of mud-filled cracks in the underlying ice. The mud has the appearance of having been squirted out of the cracks by pressure, but squeezing seems improbable for the mud is frozen a foot or two below the surface. Instead, the ridges are attributed to down-melting of the surrounding ice leaving the mud filling of the crack standing in relief. The thawed mud spreads out somewhat on flat surfaces, and mud-filled cracks outcropping on a steep slope supply material for mudflows. The mud is finer than most englacial debris, and it may have been derived from pond deposits which once overlay this spot and filled cracks in the underlying ice.

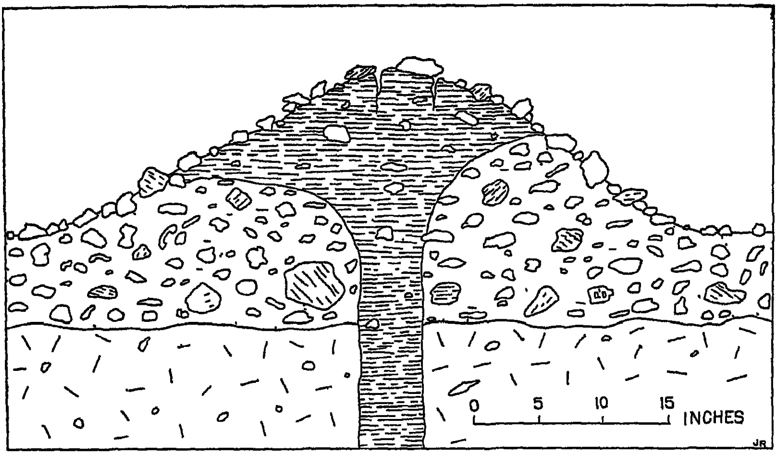

FIG. 3 Cross-section sketch of excavated mud ridge in superglacial mantle showing relation to mud-filled crack in underlying ice

FEATURES ON DEBRIS-MANTLED ICE

INFLUENCE OF DEBRIS ON SUPERGLACIAL TOPOGRAPHY

It is a matter of common knowledge that a thin covering of debris facilitates glacier wastage because the heat-absorbing fragments melt into the ice. This process has been termed indirect ablation by Philipp (1912, p. 498). A thicker mantle insulates the ice and protects it from rapid melting, and a cover of variable thickness permits different rates of ablation and produces irregularities on the glacier surface and instability in the debris mantle. An ice mound wholly or partly stripped of its debris by running water, mass movement or some other means undergoes rapid changes in size and shape by melting of the exposed ice (pl. 4, fig. 1). Complete inversions of relief occur when an ice mound, deprived of its insulation, melts down and becomes a depression. On a rough ice surface mantled with shifting debris today's depressions may well be tomorrow's high points (Russell, 1892, p. 177; Gilbert, 1904, p. 202; Sherzer, 1907, pp. 55-57; Hobbs, 1911, p. 68; Tarr and Martin, 1914, pp. 61, 206). Comparisons of Wolf Creek Glacier photographs taken in 1935, 1939, 1941 and 1947 by Walter and Foresta Wood show many changes in topography on debris-mantled ice but no major inversions. A period of several decades is probably required for a complete inversion of the heavily mantled ice features of this area. This is longer than the several seasons estimated by Russell (1897, pp. 115-117) for coastal Alaskan glaciers, but it is in accord with Sherzer's (1907, p. 57) statement that decades or centuries are required for a complete cycle in the Selkirks of Canada. Obviously many variables enter the picture including climatic factors and thickness and nature of the debris mantle.

The thickness of debris required for the development of an ice mound or ridge depends partly upon grain size, with coarser material needing greater thickness. Layers of sand ½ to ¾ inch thick, even though moist, offer sufficient insulation for the development of mounds in areas of otherwise unmantled ice, but they do not prevent melting entirely as shown by their moist condition. Dry sand is an even better insulator (Birch *et al.*, 1942, p. 259). A one-inch cover of moist sand and silt on cones of annual river ice in this area was not sufficient to prevent their complete destruction before summer's end. However, the insulating power of such a cover is demonstrated by the rapid melting which sets in when a part is removed (pl. 4, fig. 1). Mantles of coarse debris must be a foot or two thick to provide equivalent protection. The inferiority of coarse materials must be due in part to the ease with which air and water circulate through the large openings between fragments and gain access to the underlying ice. Rain water trickling down through such openings would cause considerable melting of the underlying ice even though the debris mantle were thick. On coastal Alaskan glaciers Tarr and Martin (1914, pp. 206-207) found that 5 to 10 feet of relatively coarse debris was required to produce enough stability to permit trees to take root. With thicknesses of only 2 or 3 feet, the mantle was too active and unstable for trees to grow. These figures may not be entirely representative as in the summer of 1948 alders 3 inches in diameter were observed growing on stagnant ice of the Malaspina Glacier mantled with a cover of gravel only 13 inches thick.

DEBRIS-COVERED ICE MOUNDS AND RIDGES

Debris-mantled ridges, mounds and cones of ice are abundant on the stagnant parts of Wolf Creek glaciers (pl. 3, fig. 2) and in those areas of active ice where superglacial debris is abundant and ablation great. Similar forms are described from many parts of the world (Forbes, 1859, p. 241; Tyndall, 1874, p. 116; Russell 1892, p. 177; 1893, p. 228; 1897, pp. 13, 115-117; Gilbert, 1904, pp. 198-202; Reid 1892, p. 34; Tarr and Martin, 1914, pp. 61, 205-209; Ray, 1935, pp. 319-322; Sherzer, 1907, pp. 55-57; Hobbs, 1911, p. 168; Workman, 1913, pp. 73-75; Rickmers, 1913, p. 255; Lewis, 1940, pp. 16-26).

The Wolf Creek ridges and hillocks average 40 to 50 feet high with extremes of 10 to 150 feet. The steepest debris-covered slopes are about 30 degrees where the mantle is dry but may approach 45 degrees where the mantle is moist and fine-grained, as also noted by Lewis (1940, p. 16) and Ray (1935, p. 320). Irregularities are the product of inhomogeneities in composition and distribution of the superglacial debris and of degradational processes at work on the glacier's surface. The few truly symmetrical cones (pl. 2, fig. 1) and ridges in this area are mantled largely by water-sorted sandy gravel, and their form reflects the homogeneity and mobility of this debris as well as the initial localization of the accumulation.

A debris-mantled feature develops and grows because the surrounding ice is lowered more rapidly by ablation. The higher a feature becomes the more thinly spread is its debris mantle, and simultaneously the mantle on the low areas becomes thicker by additions from surrounding areas and to some extent by residual accumulation from the melting ice. Obviously this produces a progressive decrease in differential ablation, and the height to which a mound or ridge can grow is determined primarily by the thickness and areal extent of the original debris accumulation. In the Wolf Creek area this limiting height is about 150 feet, but it will presumably be different in other areas owing to variations in factors controlling ablation.

LATERAL AND MEDIAL MORAINES ON ICE

It has long been recognized that lateral and medial moraines mark the outcrops of steep, debris-rich zones in glaciers (Salisbury, 1896, p. 800), and restatements of this principle (Washburn, 1935, pp. 1883-1884) hardly seem necessary. The debris composing such moraines is largely a residual accumulation of englacial detritus left on the surface by melting; only a limited amount has always been superglacial. In the ablation zone, moraines rise above the general surface as debris-mantled ice ridges, which become higher and broader with continued melting because of an increase in the amount of superglacial debris and its wider distribution by running water and mass movement.

Contortions in some medial moraines have been ascribed to deformation during passage over ice falls (Washburn, 1935, p. 1882), but on Wolf Creek the ability of a medial moraine to reestablish itself almost unchanged below an ice fall is demonstrated on several glaciers. In these instances, the surficial debris of the moraine collects in transverse crevasses formed at the ice fall. Ice ridges between crevasses are cleaned of their surficial debris by mass movements and calving, and subsequent rapid melting soon brings them to a level below the debris filling of the crevasses. Part of the debris then spills back over the unmantled ice, and surficial continuity of the moraine is reestablished. The only noticeable change is a somewhat greater width produced by lateral spreading. Detritus lost in the deeper parts of the crevasses is replaced by the accelerated melting of shattered debris-containing ice. Medial moraines on Wolf Creek glaciers do show contortions, however, which are related to vigorous surge-like advances of tributary glaciers as suggested by Washburn (1935, p. 1882) and to adjustments within glaciers by lateral obstructed flowage controlled by both pressure and gravity.

TRANSVERSE MORAINES

Transverse moraines are debris-covered ice ridges on a glacier's surface joining a medial and a lateral or two medial moraines. They have been described (Chamberlin, 1895, pp. 204-205; Salisbury, 1896, p. 800; Chamberlin and Salisbury, 1909, pp. 296-297, Tarr, 1908, p. 87; Tarr and Martin, 1914; p. 61, Palmer, 1924, pp. 439-440; Visser, 1932, pp. 35-36; Bretz, 1935, p. 179; Klebelsberg, 1938, p. 33; 1939, p 311) under the names cross, oblique, horseshoe, melt, and pseudoterminal moraines. Since these names are either inappropriate or have been used to designate other features, the term *transverse moraine* is suggested as most closely in keeping with the designations lateral and medial already in use. Transverse moraines are formed of debris accumulated at the end of an ice stream superimposed upon or inset within another (Workman, 1913, p. 72-73; Visser, 1932, pp. 33-39; Klebelsberg, 1938, pp. 25-31, 33). Moraines of this type have also been ascribed to crowding and pinching out of small ice streams by a more powerful glacier (Palmer, 1924, pp. 439-440; Washburn, 1936, p. 488), but the best example on Wolf Creek was at the end of an inset ice stream.

A closely similar feature is formed along the outcrop trace of parabolic debris bands in glaciers These are probably shear zones along which debris has been carried up into the glacier from its base. François E. Matthes[2] has suggested that the debris-covered ice ridges marking the trace of such bands might be termed *transverse moraine bands.* Those seen on Wolf Creek extend entirely across a glacier from one lateral moraine to the other and are clearly concordant with the ribboned structure and thinner debris bands in the ice.

[2] Personal communication

FORBES DIRT BANDS

Forbes dirt bands appear as a series of parallel, parabolic bands of debris on the surfaces of some valley glaciers below ice falls. They are rare, wholly surficial, and not to be confused with debris bands within the ice. Since their early description (Forbes, 1859, pp. 25, 39-41, 213-214), they have attracted attention of glaciologists and explorers (Tyndall, 1874, pp. 130-131; Sherzer, 1907, pp. 39, 50-54; Washburn, 1935, pp. 1885-1889). Fisher (1942; 1944, pp. 26-40) presents a recent study of this phenomenon. Forbes dirt bands are seemingly related to great glacier-wide crevasses developed during periodic, possibly annual, movements of valley glaciers over ice falls. Some workers hold that rapid melting of shattered ice in the crevasses during summer forms a series of broad depressions in which debris collects from the melting ice (Washburn, 1935, p. 1889) or other sources (Sherzer, 1907, pp. 50-54). Fisher (1942, pp. 8-9; 1944, pp. 30-36; 1947, pp. 139-140) concludes, however, that the crevassed areas constitute insets of relatively clean ice, so the dirt bands must be parts of the original, somewhat dirty glacier surface preserved on blocks between crevasses.

In spite of special search no Forbes dirt bands were found on Wolf Creek glaciers, although the numerous ice falls and vigorous ablation in this area would seem to favor their development. The debris of medial and lateral moraines passing over ice falls collects in transverse crevasses, as described above, and this produces local concentrations of detritus resembling segments of dirt bands. These are termed "inverted" Forbes dirt bands by Fisher (1942, p. 12), but they are short lived as they disappear when the moraines are re-established in the manner previously described.

GLACIAL TABLES

Glacial tables are isolated rock fragments perched on ice pedestals. They form because of the protection from melting afforded by the rock and are most numerous in areas covered by a sparse superglacial mantle containing large fragments. The smallest Wolf Creek tables consist of stones 6 inches to 1 foot in diameter perched on pedestals less than an inch high. The largest involve fragments several feet in diameter on pedestals as much as 4 feet high. Most have the southward tilt reported in other parts of the northern hemisphere (Russell, 1893, p. 227; 1897, pp. 11-12; Sherzer, 1907, pp. 56-57) that is caused by greater melting on the south side, probably by reflected radiation (fig. 4). Here, the amount of tilt does not exceed 25 degrees, and the slope of the ice surface and shape of the rock fragment occasionally outweigh sun position to cause tilts in other directions. The numerous unseated tables, and the lack of pedestals higher than 4 feet are attributed to growing instability of perched boulders as the height of pedestal and degree of tilt increase. It seems that average heights of pedestals in high latitudes must be less than in low latitudes for this reason. Pedestal-type nieve penitente (Workman, 1914, pp. 320-323), formed by dislodging of perched boulders, are also present.

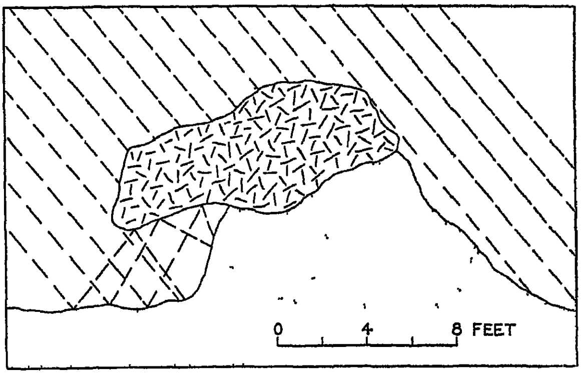

Fig 4. Sketch showing how reflected radiation causes unequal melting and southward tilt of glacial tables in the Northern Hemisphere.

Size of rock fragments is an important factor in the formation of glacial tables (Hobbs, 1911, p. 167; Ray, 1935, p. 322; Tollmer, 1938, pp. 44-46), and the minimum size required increases as the season becomes warmer (Fryxell, 1933a, pp. 643-646). Color, lithology, shape, and position of the fragment are also significant. On one Wolf Creek glacier a slab of white marble 2 inches thick and 10 inches across had formed a table by mid-July while a nearby diabase of similar width but fully 10 inches thick had melted into the ice. In this case color and thermal conductivity were significant with color probably outweighing conductivity (Birch *et al.*, 1942, pp. 253, 255-256). Conductivity may have greater influence in other instances as it differs considerably among rocks. The point, that flat slabby pieces of rock casting broad shadows form tables where smaller fragments of similar thickness do not (Hobbs, 1911, fig. 92, p. 167), was confirmed.

By mid-August glacial tables, particularly those formed by smaller stones, were seen only in the higher parts of the ablation zone. This indicates that air temperatures must be such that melting of shaded ice by conduction and condensation from the air is measurably slower than melting of unshaded ice by conduction, condensation, and radiation. The zone most favorable for development of glacial tables moves gradually up a glacier as the summer season progresses. It may also descend in fall, although the descent is probably cut short by snow fall.

Glacial tables are significant beyond the mere fact of their existence. They provide a minimum measure of ablation, but determination of the time interval involved is difficult except by direct observation. In this respect Russell's (1893, p. 227) report that tables on Malaspina Glacier seldom survive the winter is intriguing, for tables on such a glacier would afford a measure of seasonal ablation up to the time of dislodgement; and if any survived the entire summer, they would give a minimum measure of annual ablation. Russell's observation needs confirmation, however, for the means by which the stones could be dislodged in winter is not apparent. In many areas stones are dislodged from their pedestals several times a year (Fryxell, 1933b, p. 742; Tollmer, 1938, pp. 44-46), so the height of pedestal measures but a fraction of the annual ablation.

Glacial tables are further noteworthy for providing a mechanism by which superglacial stones can be shifted about independently of glacier movement, as demonstrated by Fryxell (1933b). This is brought about by the sliding and rolling of stones dislodged from ice pedestals. The direction of movement is controlled by differential melting, slope of the glacier's surface, and shape of the stone, and it may be at variance with the direction of glacier flow. The movement may also be of greater amount, than by flowage, and stones on the Teton Glacier, Wyoming, moved an average of 24.5 feet per month during summer by this process (Fryxell, 1933b, p. 742). On most Wolf Creek glaciers the direction of slope is opposed to the direction of tilt of many perched stones so little movement results when they slip off their pedestals.

DUST WELLS AND DUST BASINS

Small holes or depressions produced by localized melting under thin accumulations of dark silt in areas of relatively clean glacier ice are known as dust wells and dust basins (Hobbs, 1911, pp. 168-169) or *Kryokonitlocher* (Nordenskjold, 1870). Brandt (1932, p. 86) distinguishes 6 types of Kryokonitlocher, but only two of them, the polar and the alpine types are recognized on Wolf Creek.

Description.—Dust wells, the polar-type *Kryokonitlocher* (Steinbock, 1936, pp. 6, 19), are small vertical cylindrical holes filled nearly to the brim with water and containing a thin layer of dark bottom silt (pl. 4, fig. 2). In places Wolf Creek Glacier is literally perforated with dust wells ¼ to 3 inches in diameter and 4 to 8 inches deep. Depths up to 24 inches have been reported from north Greenland (Philipp, 1912, p. 489), and the large-diameter wells are generally the deeper (Poser, 1933, p. 5). Small wells are circular, but large ones tend to be oval, oblong, or crudely scalloped.

An accompanying form is the larger semicircular tub or basin also filled with water and containing a layer of dark bottom silt, 1/16 to ⅛ inch thick. This is the dust basin or alpine-type *Kryokonitlocher* supposedly predominant on temperature glaciers (Steinbock, 1936, pp. 6, 19). Elongated basins are baignoire[3] and those in which the elongation is determined by sun position are mid-day *Kryokonitlocher* (Wagner, 1938, pp. 130-131) or *trous meridiens* (Agassiz, 1847, p. 101).

The largest dust basin observed was 3 feet long, 1½ feet wide, and 12 to 16 inches deep, but most Wolf Creek basins are about half that size and some are scarcely larger than big dust wells. Wagner (1938, p. 130) reports basins of similar size but greater depth, 30 inches, in north Greenland. Semicircular outlines predominate, and larger basins have a consistent east-west elongation, as noted by Agassiz (1847, pp. 102-103), that is independent of surface slope and structures within the glacier. It parallels the banded structure of the ice only in those places where the bands are east-west.

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[3] Hobbs (1911, p 168) spells the word "bagnoire" although the French word baignoire, meaning bath tub, is probably intended as that is the spelling used by Agassiz (1847, p 100).

The north wall of Wolf Creek dust basins is smoother and more gently inclined than the nearly vertical south wall (fig. 5), which is irregularly indented by remnants of small dust wells. The gentle but prevailing northward slope of the basin floor is interrupted at the foot of the south wall by a low step or sill an inch or two high. The upper surface of this sill, only partly mantled with silt, is pitted with shallow dust wells. Other parts of the basin floor relatively free of silt also project above the average level, and small pebbles on the floor may be perched on low pedestals. Basins are generally filled with water to the level of the lowest outlet.

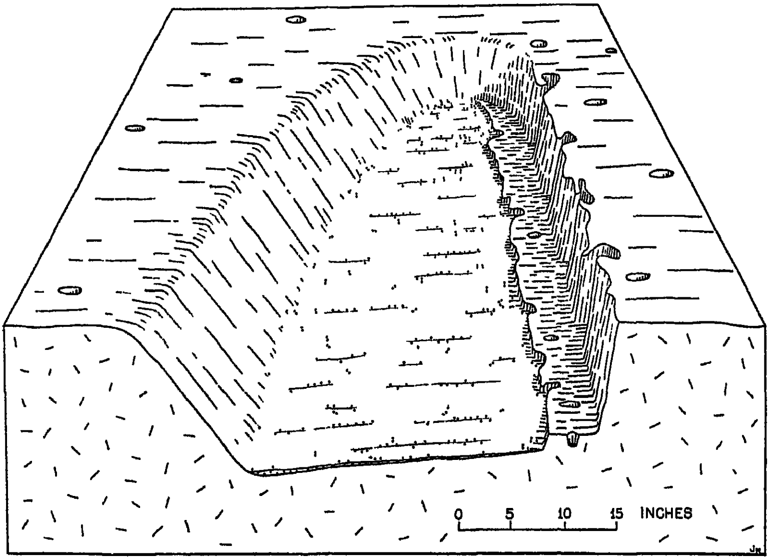

Fig 5 Diagrammatic sketch of a dust basin showing typical asymmetrical north-south cross section, layer of dark bottom silt, sill at base of north wall, and irregularities of the nearly vertical north wall.,

Where dust wells and basins are abundant, their distribution and arrangement, though not shape, are concordant with the banded structure of the ice. They are most numerous in bands of white porous ice and almost entirely lacking in clear blue ice. Neither was found in ice heavily mantled with superglacial debris nor in the lower parts of the ablation zone.

Only one *Kryokonit* pool (Poser, 1933, p. 6; Wagner, 1938, p. 130; Steinböck, 1936, p. 8) was seen. This depression, 60 feet long by 30 feet wide and 12 to 15 feet deep, contained water about 4 feet deep and had a thin irregular mantle or dark bottom silt.

Origin.—Dust wells and basins are briefly noted in the English literature (Tarr, 1897, p. 141; Barton, 1897, p. 216; Sherzer, 1907, pp. 54-55; Chamberlin and Salisbury, 1909, pp. 269-271; Hobbs, 1911, pp. 167-169; Workman, 1914, p. 312; Wright and Priestly, 1922, pp. 283-286; Odell, 1925, pp. 305-306, Thwaites, 1946, p. 5), but German and Scandinavian writers provide a more detailed analysis of their origin.

Superior development of wells in northernmost latitudes and of basins in more temperate areas (Steinbock, 1936, pp. 6, 11, 15, 19) forms the basis for designating wells as the polar-type and basins as the alpine-type *Kryokonitlöcher.* However, neither type need be wholly excluded from either environment (Sherzer, 1907, pp. 54-55; Odell, 1925, p. 305; Wagner, 1938, pp. 130-131; Brandt, 1931a, p. 125), and this separation does not apply on Wolf Creek where dust wells and basins are closely associated. The fine debris in these depressions is commonly called dust, and some writers specifically speak of it as windblown (Chamberlin and Salisbury, 1909, pp. 269-271; Philipp, 1912, pp. 497, 504-505; Poser, 1933, p. 8; Steinböck, 1936, p. 19; Brandt, 1931b, p. 320). This practice is partly due to the fact that these forms were early described from the Greenland ice sheet where such fine debris is probably dust. Nordenskjold's (Steinbock, 1936, p. 2) postulate that some of the dust is of cosmic origin has not been confirmed by chemical analyses (Poser, 1933, p. 8).

A small part of the silt in the Wolf Creek depressions may be of eolian origin, but most of it is probably fine englacial debris melted out of the ice and accumulated in small depressions on the surface by films of melt water. The material composing small mud balls on Alaskan glaciers (Ray, 1935, pp. 318-319) is presumably of this origin. In some instances the work of melt-water sheets and films in transporting the silt may be secondary and supplementary to initial distribution by wind (Workman, 1914, p. 312) or it may be primary, as thought likely on Wolf Creek, in which case the terms dust well and dust basin are hardly appropriate. *Kryokonitlöcher* is even less appropriate since *Kryokonit* is supposedly dust of cosmic origin (Steinböck, 1936, p. 2). However, the terms dust well and dust basin have become so firmly established by long usage that they are used here regardless of the origin of the bottom silt.

The initial depressions in which silt accumulates can be formed by intergranular melting or by localized melting of exceptionally porous air-filled ice. The initial deposit is augmented by additional silt washed into the depression or by merging with neighboring accumulations.

It is agreed that these thin accumulations cause melting by absorption of direct solar radiation, but melting by this means ceases when the depth of the well is such that its bottom is continually shaded. Many wells exceed this depth, and various explanations for the over-deepening have been advanced. Drygalski (Steinböck, 1936, p. 7) favors deepening by warmed surface water which sinks to the bottom of the hole because of its greater density at 4°C. than at 0°C. Philipp (1912, pp. 490-491) suggests that deepening is produced by radiations transmitted obliquely through the ice to the silt. Wagner (1938, pp. 133-137) also favors transmitted radiations, and Poser (1933, p. 9) testifies that water and ice surrounding the silt are permeable to such radiations, which of course may be both direct and diffused. If the bottom silt is warmed by transmitted radiations, the sinking of warmed surface water to the bottom becomes less significant because the bottom water would also be warmed and correspondingly heavy.

Drygalski (Steinbock, 1936, p. 9) is said to have demonstrated that deepening of wells ceases when the surface of the water filling them freezes but is resumed when the ice layer melts. This is advanced as an argument against deepening by transmitted radiations on the basis that such deepening should continue regardless of an ice cover on the water. This argument is weakened by the possibility that the conditions causing freezing could be related to or accompanied by a decrease in radiation.

Steinböck (1936, pp. 8-9) suggests that the efficiency of transmitted rays decreases with increasing latitude owing to greater reflection from the ice surface as the sun's elevation becomes lower. He also holds that diffused radiations from the sky reach the bottom of wells directly, but Wagner (1938, p.

132) questions the efficacy of this process and maintains that only 1 per cent of the total radiated energy can reach the bottom of the average well as diffused radiation. This would scarcely be enough to produce melting at a rate greater than the lowering of the ice surface by direct ablation. However, Wagner may have underestimated the amount of diffused radiation which reaches the bottom of a well, for on clear summer days diffused heat rays constitute as much as 1/8 to 1/3 of the total solar radiation (Lundegårdh, 1931, p. 4), and the amount of diffused radiation increases directly as the cloud cover to approximately 100 per cent of the total radiant energy. Wright and Priestly (1922, p. 5) also note that at very low solar altitudes the "sky radiation" on a horizontal plane is more intense than direct solar radiation. Thus, at high latitudes and in cloudy weather diffused radiation may play a much larger role than imagined by Wagner, and certainly some of the energy transmitted through the ice arises from diffused radiation.

The possible effect of radiations reflected to the bottom of the hole from its walls is difficult to evaluate but is thought to be an additional minor factor. Steinbock (1936, pp. 10-11) reports that organisms living in the bottom mud (Sherzer, 1907, pp. 45-55) generate some heat, and this may be a further means of deepening the wells.

Final conclusions on the relative significance of these various factors do not yet appear possible, but it would seem that if one imagines himself at the bottom of a dust well he would see more light, and therefore more energy, coming in as diffused radiation at the top of the well than as transmitted radiation through the walls. For this reason it is suggested that of all the above mentioned factors diffused radiation is probably the most effective in deepening the dust wells after the bottom is shaded from direct radiation.

Dust basins are of the same genus as dust wells, but the relative efficacy of genetic agents is different. Some basins may be formed by melting under initial accumulations of silt or through the gathering of such debris into pre-existing shallow depressions (Poser, 1933, p. 17). Others are developed by integration of adjacent dust wells (Hobbs, 1911, p. 167; Philipp, 1912, p. 497; Brandt, 1932, p. 93). Direct radiation has a relatively large part in forming basins by virtue of ready access to the floors and walls. In northern latitudes lateral enlargement takes place chiefly through melting of the north, east, and west walls by direct radiation. The north wall becomes especially smooth and acquires a slope about normal to the sun's rays (Wagner, 1938, p. 136). In northern latitudes the south wall and the basin floor at its foot are in shadow. This accounts for the irregularities of the south wall and adjacent basin floor, for melting there must be by transmitted, diffused, and reflected radiations or by convection and conduction in the air and water. Most irregular indentations of the south wall and small depressions on the basin floor are remnants of incompletely digested dust wells overtaken by the retreating walls, although some may be a later development.

Northward slope of the basin floor results from greater exposure to direct radiation on the northern part of the floor. Pebbles perched on small pedestals on the floors of some dust basins give further proof that much of the bottom melting is by direct radiation. The lip of the sill at the base of the south wall marks the edge of the mid-day shadow, and the height of this sill, an inch or two, measures the superiority of melting by direct radiation over melting by all other means.

The consistent east-west elongation of dust basins on Wolf Creek, in the Alps (Agassiz, 1847, pp. 102-103), and in the Himalayas (Workman, 1914, p. 312) is related only locally to visible structures in the ice. The elongation on Wolf Creek is ascribed to the greater amount of radiated energy received by the east and west walls together than received by the north wall alone, despite the recognized efficacy of the midday sun. Combined retreat of the east and west walls is therefore greater and an east-west elongation results.

According to Wagner (1938, pp. 134, 137) dust basins are developed largely in porous air-filled ice because it is a poor conductor of radiations, and they are converted to heat. Contrariwise, dust wells are supposedly formed in clear ice because it is a better conductor, and the radiations are transmitted to the bottom silt. This thesis is not confirmed on Wolf Creek as both depressions favor porous ice, and Ahlmann's (1935, p. 104; 1938, p. 417) observation that ice absorbs more radiation than snow suggests that Wagner's conclusions may not be universally applicable. An alternate possibility is that melt water washes fine debris off the smooth surface of clear ice before it has time to become firmly seated by melting. This might not occur if the temperature were always below freezing and no melt water were available, and the nature of the ice in which dust holes form need not be the same everywhere.

The localized melting that produces wells and basins is called indirect ablation (Philipp, 1912, p. 498) or *Kryokonitablation* (Poser, 1933, p. 10). Observations and experiments (Brandt, 1931a, p. 126; 1932, pp. 88-91) show that clear cold weather with much radiation favors development of wells and basins by allowing indirect ablation to proceed at a faster rate than direct ablation. Warm cloudy weather produces more surface melting by convection, conduction, and condensation and is less favorable to the development of dust holes. If direct ablation becomes superior, it can eventually destroy the dust holes. The maximum depth attained by these features is controlled by the relative rates of direct and indirect ablation (Brandt, 1931a, p. 126), and in some areas wells and basins are deeper in spring and early summer than later in the season when direct ablation becomes greater (Poser, 1933, p. 10). In late summer, dust holes on Wolf Creek glaciers were limited to a narrow zone lying just below the firn line. Their absence from lower parts of the ablation zone is attributed to the superiority of direct ablation there. With advance of the summer season, the zone of dust wells and basins probably moves slowly up the glaciers, following the retreating snowline.

Dust wells and basins are instrumental in causing melting at air temperatures far below freezing (Philipp, 1912, pp. 489-490; Workman, 1914, pp. 313-314). They may be a major factor in glacier ablation in high latitudes and at high altitudes (Hobbs, 1944, p. 82).

Steinböck (1936, pp. 9-11, 20) suggests that the relative effectiveness of direct radiation in middle latitudes and of diffused radiation farther north accounts for the development of polar- and alpine-types of *Kryokonitlocher*. Wagner (1938, p. 137) believes it is more a matter of ice condition, with clear ice in the north favoring development of wells and air-filled ice in the south favoring basins. Insofar as basins are formed through integration of wells, they are both the result of the same processes, although direct radiations are of more significance in developing basins and diffused and transmitted radiations of greater effect in forming wells. Integration of wells to form basins probably occurs less easily in areas where the air temperature is continuously below freezing than in warmer regions because back-melting of the walls by conduction is less. Thus, air temperature may be as fundamental as any other factor in accounting for the abundance of wells and scarcity of basins in far northern latitudes. It is possible that wells in regions like Wolf Creek are formed chiefly in the spring and fall, and basins develop largely in mid-summer when air temperature rises above freezing for longer periods.

In summary, dust wells and dust basins are found on relatively clean parts of Wolf Creek glaciers where indirect ablation exceeds direct ablation. In this area much of the bottom silt need not be dust. Over-deepening of dust wells seems to be due largely to diffused and transmitted radiations, and dust basins are more largely the product of direct solar radiation. East-west elongation of the basins is independent of ice structures and is ascribed to the greater combined melting of the east and west walls than of the north wall alone. The separation of alpine- and polar-type *Kryokonitlocher* does not hold on Wolf Creek.

RELATION TO GLACIAL DEPOSITS

The transition from superglacial accumulation to glacial deposition is accomplished in the final stage of glacier degradation where the ice thins, stagnates, and ultimately disappears. All superglacial materials and features described above are subject to alteration as long as they are underlain by ice, and many, such as glacial tables, dust wells and dust basins, will be destroyed. The amount of modification to be expected during this transition decreases as the ratio of debris to ice increases. In the final stage the debris mantle may be so thick and melting so slow that the melt water filters off gradually without reworking the detritus. However, the space vacated by melting ice must be filled, so some modification of superglacial forms is inevitable. Large depressions extending through the glacier to its floor are rare, and Cook's (1946, pp. 575-577) perforation mechanism appears to be based on improbable assumptions (Holmes, 1947, p. 244). In the Wolf Creek area glacier ponds enlarge themselves by lateral undercutting and melting of the ice walls (pl. 3, fig. 1), but there is every indication that rapid accumulation of bottom debris prevents deepening by melting except possibly in the initial stage.

Complete wastage of stagnant ice heavily mantled with superglacial debris leaves an accumulation having the constitution and topographic form of the superglacial mantle. However, the topographic relief will be on a smaller scale, for it can be no greater than the thickness of the superglacial mantle. In constitution the material may range from well-sorted and well-bedded clay, silt, sand, and gravel to an ill-sorted mixture of large angular blocks. In general, it will be distinguished by its loose texture, lack of clay and silt, except in pond deposits, and evidences of reworking by running water.

In the instance of valley glaciers, streams of melt water coursing over the valley floor usually rework and largely destroy such accumulations, although they may be preserved in protected spots. Areas of continental glaciation afford better opportunity for preserving accumulations of superglacial origin.

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THERMODYNAMIC RELATIONS OF IMMISCIBILITY AND CRYSTALLIZATION OF MOLTEN SILICATES

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ABSTRACT. The relations of immiscibility and of crystallization in silicate melts and magmas are discussed On the basis of the ionic constitution of the melts it can be said that regions of immiscibility may exist at composition, temperature, and pressure ranges which have not been explored in the laboratory.

During fractional crystallization of a silicate system the regular progressive precipitation of the cations and of the anions is governed by the relations of the E-values (activation energies of migration), and of the pO-values (oxygen ion concentrations) respectively.

INTRODUCTION

MODERN igneous geology has developed in consequence of a rapidly increasing knowledge of the crystallization processes of silicate melts. But thus far little has been known about the physicochemical properties of the melts, and practically nothing has been applied to geological problems. In recent years pertinent experimental data have accrued from many sources, and it now would seem possible to discuss some of the results of direct bearing on petrology.

In the constitution of silicate melts lies the key to the problem of the interesting appearance of liquid immiscibility and to the nature of the solid phases which may crystallize out. Liquid immiscibility was first discovered in silicates by Greig (1927) and has aroused much discussion, but the consensus of opinion has been that no geological significance could be attached to it. However, the discussion now has been revived by Fenner (1948), whose paper on immiscibility of igneous magmas was published while the present manuscript was in preparation. Fenner maintains that immiscibility is thermodynamically probable, discusses the petrological implications, and presents field evidence for its occurrence in natural magma.

In the present paper certain properties of artificial melts only will be discussed, and no attempt will be made to survey the applications to geology.

CONSTITUTION OF SILICATE MELTS

The entropy of fusion of silicates is not high, so that no major changes in structure are to be expected on melting.

On general grounds one would, therefore, expect the liquid structure to resemble the solid. Experiments have confirmed this· see e.g. Blitz (1934), Biscoe et al. (1938), Endell (1942), Martin (1943), Zachariasen (1932).

Molten silicates possess an *ionic* constitution in analogy with solid silicates. The extent of ionic dissociation is comparable to that of aqueous solutions of strong electrolytes.

Typical cations are Na^+, K^+, Ca^{2+}, Mg^{2+}, Fe^{2+} which occupy no fixed positions in the melt but exhibit a high degree of mobility.

The *anions are predominantly polymerized silicon-oxygen* tetrahedra linked together by oxygen bridges to one-, two- or three-dimensional networks. The network is similar to that found in crystalline silicates but more irregular.

The bonds between the central silicon ion and the oxygens at the corners of the tetrahedra are mainly covalent, but may also be regarded as a largely polarized ionic link (which means the same).

A verification of the existence of a polymerized silica-network is given by the high viscosity of silicate melts. The viscosity is found also to increase upon adding Al_2O_3. This shows that Al can replace Si in the polymerized anion, as it does in some crystalline silicates, e g., feldspars. To a certain extent three-valent iron forms anions, probably of the form $(FeO_6)^{9-}$.

In addition to the anions mentioned above, the melt contains small amounts of free oxygen ions, 0^{2-}, the amount of which is determined by the following reactions:

$$
\begin{aligned}
(SiO_4)^{4-} &\rightleftharpoons (SiO_{3.5})^{3-} + \tfrac{1}{2}\,O^{2-}\\
(SiO_{3.5})^{3-} &\rightleftharpoons (SiO_3)^{2-} + \tfrac{1}{2}\,O^{2-}\\
(SiO_3)^{2-} &\rightleftharpoons (SiO_{2.5})^{1-} + \tfrac{1}{2}\,O^{2-}\\
(SiO_{2.5})^{1-} &\rightleftharpoons (SiO_2) + \tfrac{1}{2}\,O^{2-}
\end{aligned}
$$

For each reaction we can write the corresponding equilibrium equation, for example:

$$\frac{[SiO_{2.5}]\cdot[O^-]^{1/2}}{[SiO_3]} = K \;\ldots\ldots\;\;\ldots\ldots\ldots (1)$$

where the square brackets indicate the activity of the components, and K is the equilibrium constant.

The activity of the oxygen ion may be taken proportional to its concentration, and it is readily seen that a large oxygen ion concentration pushes the reactions to the left, i.e., the oxygen ion concentration increases with increasing basicity of the melt. It becomes appreciable in melts of more basic composition than that corresponding to an orthosilicate. Thus the so-called "pO-value" is a measure of the basicity of the silicate melt.

In formal analogy with the symbol pH, the oxygen ion activity is indicated by its negative logarithm and denoted by the symbol pO.

The degree of polymerization is markedly influenced by the nature of the cations present. This means that the nature of the cation influences the activity coefficient of the different types of macroions. The effect is that at constant basicity the formation of oxygen bridges is favored by a weakly polarizing ion, as K^+, and decreases in the sequence K^+-Na^+-Li^+. Flood and Forland (1947) have shown that at a certain basicity where the K silicates contain anions with about *2.5* oxygen/Si, the Na silicates have *3* oxygen/Si, and the Li silicates *4* oxygen/Si. This is in agreement with the fact that in the alkali silicate melts the viscosity decreases in the sequence $K > Na > Li$, i.e., with increasing polarizing power of the cation. See e.g. the compilations by Eitel (1941).

It is probable that in silicate melts containing earth-alkali and ferrous ions, the tendency is even greater towards a breaking up of the macroions into smaller units. This is indicated also by the fact that the viscosity of iron silicate melts is rather low even up to high SiO_2 content.

When CO_2 is dissolved in molten silicates it reacts with the oxygen ions to form CO_3^{2-} ions.

$$CO_2 + O^{2-} \rightarrow CO_3^{2-}$$

At constant CO_2 pressure the amount of CO_2 dissolved (the concentration of CO_3^{2-}) will be approximately proportional to the oxygen ion activity of the melt. This is the method used by Flood to determine oxygen ion activities.

When H_2O is dissolved in the melt the following reaction takes place

$$H_2O + O^{2-} \rightleftharpoons 2(OH)^-,$$

in harmony with the fact that $(OH)^-$ is an individual building brick in the crystal structure of many silicate minerals.

LIQUID IMMISCIBILITY

Silica melts and melts of Na_2O, K_2O or Al_2O_3 are known to be completely miscible in the low-temperature region. But silica shows only partial miscibility with melts of CaO, MgO, ZnO, FeO, Fe_2O_3: at high contents of SiO_2 an area of immiscibility exists. The effect of the various oxides is of considerable interest. The elements of the vertical column of the periodic table Mg, Ca, Sr, Ba, exercise, with increasing basicity, a successively decreasing effect; next follow the alkali metals Li, Na, K, Rb, Cs. The shape of the various melting curves is interesting: not until Rb and Cs is the normal shape restored. See Kracek (1930).

Strongly polarizing cations produce two effects· They raise the activity coefficient of highly polymerized silicon-oxygen networks, and favor the formation of individual (SiO_4)-groups. Under certain conditions melts high in (SiO_4)-groups will, therefore, unmix from melts of highly polymerized silicon-oxygen networks. These two melts will be referred to as ortho-melts and para-melts, respectively.

In various ternary systems of rock-making oxides a region of immiscibility appears. Again highly polymerized para-melts separate from melts rich in (SiO_4)-anions. But no complete investigation of any binary or ternary silicate system has been attempted at temperatures somewhat above those of crystallization, nor is the effect of pressure on the solubility gap known.

Our ignorance is deplorable, for a complete knowledge of the solubility relations would be of the greatest value for the theory of magmatic differentiation. Since the experimental data are incomplete, it is impossible at present to predict anything regarding the general immiscibility relations in silicate systems. However, it can be said that regions of immiscibility may exist at composition, temperature, and pressure ranges which have not been explored. It is not permissible, therefore, on the basis of present knowledge, to conclude that differentiation by magmatic immiscibility is impossible or unlikely. For all we know, it may be widespread.

The preceding conclusions are supported in a general way by Buchner (1906) and by Findlay (1935) [See also Marsh (1935) and Hildebrand (1936)].

According to the general discussion presented by Findlay the solubility curve of two partially miscible solutions may show either an upper or a lower critical solution temperature, above or below which complete miscibility exists. Investigations of a number of systems containing organic liquids indicate that the general form of the concentration-temperature curve is that of a closed ring. Theoretically this ring may be cut off by the stability fields of crystalline phases, as actually observed in dry silicate melts where the lower portion of the hypothetical ring is dissected. Experiments show that by pressure the lower critical solution temperature is always raised, the upper may be either raised or lowered.[2] According to a recent review in Chem. and Eng. News, Sept. 1948 (p. 2877) a lower critical solution point has been found by Secoy in the system UO_2SO_4-H_2O under pressure.

The idea presenting itself on the basis of the facts just mentioned is that certain silicate systems may possess regions of immiscibility at higher temperatures although no immiscibility has been encountered close to the temperatures of crystallization. It is furthermore conceivable that under pressure the solubility gap in, say, the system CaO-SiO_2 would disappear *at the temperature of crystallization*, but exist in the shape of a closed ring at elevated temperatures. An analogy to this may exist in the system BaO-SiO_2 at ordinary pressure. the shape of the crystallization curve here indicates that an area of immiscibility exist in the neighborhood, perhaps as indicated in figure 1.

In a system of two partially miscible liquids the addition of a foreign substance will expand the region of immiscibility if the solubility of the foreign substance is markedly different in the two liquids. In view of this fact the effect of carbon dioxide and water (and fluorine) may be considered.

Water and carbon dioxide are more readily dissolved in ortho-melts than in para-melts; therefore, the immiscibility relations will be affected and the solubility gaps may be widened by the addition.

It has been shown by Tuttle and Friedman (1948) that the system Na_2O-SiO_2, which is miscible in all proportions, splits by addition of H_2O, into two immiscible liquids, one very rich in water (corresponding to an ortho-melt?), one rather rich in silicate (corresponding to a para-melt). The immiscibility region has the shape of a spheroid. The upper extent is not known, but parts of the upper portion are cut by the stability fields of $Na_2Si_2O_5$ and quartz. The lower portion is not cut off by any stability fields of crystalline phases, but shows a lower critical solution temperature between 200 and 250°C, below which complete miscibility occurs.

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[2] Theoretical considerations suggest that the immiscibility of a conjugate pair of liquids of simple molecules or atoms (e g, alloys) continues to increase with decreasing temperature But if one or more of the constituents of the liquids undergo polymerization or inter-molecular rearrangement by cooling, then a lower critical point is possible This situation is often encountered in mixtures of organic compounds, of sulfur, etc Likewise, it is to be expected in silicates

It is of more than usual interest that the immiscibility spheroid in this system possesses a lower consolute point. It demonstrates that, in this silicate system, the theoretical behavior illustrated by figure 1 has been realized.

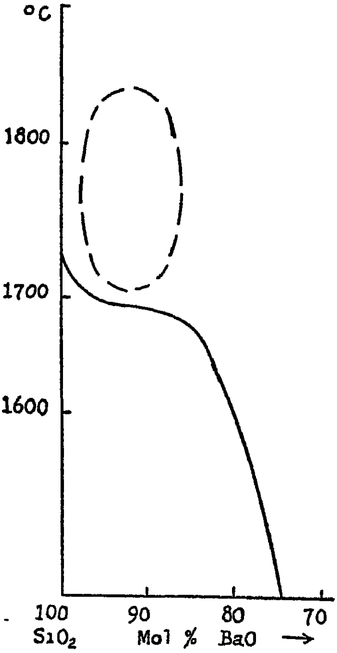

Fig 1 The liquidus curve in part of the system SiO_2-BaO. A hypothetical region of immiscibility is indicated.

This fact lends added support to the view expressed above that immiscibility in silicate melts would take place in closed regions possessing a definite downward termination, and that such regions may be present in certain silicate systems at temperatures well above crystallization.

CRYSTALLIZATION IN SILICATE MELTS

The theory of crystallization and melting may be studied from many angles. The distinction between melt and crystal may be said to consist, e.g., in the degree of order of the ions. At the point of crystallization an ordered arrangement becomes themodynamically stable, or, put into other words, for a given composition the activity of the constituent ions becomes lower in an ordered state than in the melt.

During fractional crystallization more and more ions gradually pass from the molten into the solid state. Wickman (1943) has shown that the so-called E-values (the activation energies of migration) of the various ions constitute a decisive factor in determining the sequence of crystallization of the several cations. Thus is explained why, in ordinary rock magmas, the ratios Fe:Mg, and alkali:lime increase with increasing crystal-fractionation.

The fractional crystallization of the anions may be understood from a study of their activity relations: Under the prevailing pO-conditions during the cooling of an ordinary basaltic magma two types of anions preserve a high activity in the melt, and therefore, appear in the first crystallization products, viz., (1) the orthosilicate anion, (SiO_4), and (2) the highly polymerized anion $[Si_{1/2}Al_{1/2}O_2]_n$. It is of more than ordinary interest that these two crystallization products in their ionic constitution are respectively analogous to the two immiscible liquids (ortho-melt and para-melt) as discussed above. It shows that homogeneous magma has an intrinsic tendency to break up into two parts that are either liquid or solid, the one rich in orthosilicate anions, the other rich in parasilicate anions, corresponding respectively to the beginning of the ferromagnesian (olivine) reaction series, and the feldspar (anorthite) reaction series of Bowen.

The reason for the well-known evolution of the ferromagnesian reaction series, olivine → pyroxene → hornblende → mica is given by equation (1). Early crystallization of olivine rids the magma of both (SiO_4)-anions and Mg-Fe cations; thus, the basicity decreases and anions of successively lower oxygen-silicon ratios increase their activity. Consequently, chain and sheet silicates [corresponding to pyroxene (hornblende) → mica] follow the crystallization of olivine.

The study of the crystallization of the aluminum-silicon-oxygen anions is not so far advanced. The high activity of the polymerized anorthite anion at low pO-values is not yet confirmed by experimental data. However, the reason for the development of the reaction series anorthite → albite is obviously related to the great thermal stability of three dimensional frameworks in which the ratio Si:Al is 1:1, as already pointed out by Wickman, loc. cit. With increasing fractionation (and increasing pO-values) the ratio Si:Al increases to 3; consequently, a series of successively more acid (albitic) plagioclases follow the crystallization of anorthite.

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A NEW SPECIES OF *PALEORHINUS* FROM THE TRIASSIC OF TEXAS[1]

WANN LANGSTON, JR.

ABSTRACT. A new species of paleorhinid phytosaurs, *Paleorhinus scurriensis*, from the basal Dockum of Scurry County, Texas, is described The lateral position of the quadrate foramen first reported in *P bransoni*, is corroborated and is considered characteristic of the genus The otic opening is present as a deep notch rather than as a completely bounded foramen as described in *P. bransoni*, and there is therefore more reason to suspect misinterpretation of this region in the genotype A further instance of the presence of a probable persistent parietal foramen in an immature phytosaur is noted. Youthful characters in the genus are discussed and the species is compared with other phytosaurs Observations are offered on the geologic significance and provisional limitations are placed upon Dockum-Popo Agie correlation.

INTRODUCTION

RECENT examination of the Triassic vertebrate material housed in the West Texas Museum (Texas Technological College) has revealed the presence in that collection of a number of hitherto unrecognized forms from the Dockum group of Western Texas. Among these is a skull of a small phytosaur collected in 1937 by Dr. John Clark on the farm of Mr. H. G. Bryan some two and one half miles northeast of Camp Springs, Scurry County, Texas. The horizon may be referred to the basal Dockum[2] which in this area is a fine to coarse sandstone and conglomerate which, although white in an unweathered state, appears brownish after prolonged exposure.

The skull when found was enclosed in a concretionary mass of white, fine-grained, calcareous sandstone that has served to protect the bone from rapid disintegration. The greater portion of the snout, the upper and lateral surfaces in the orbital region, and the lateral bones surrounding the temporal fenestrae are lacking. Otherwise the skull is free of distortion and the majority of the sutures are remarkably distinct. Unfortunately the palatal and posterior cranial regions have not been completely freed from the matrix and therefore these areas are only partly available for discussion.

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[1] Expenses connected with the preparation of the present paper were defrayed by the Research Council of the Texas Technological College, Lubbock, Texas

[2] Beede and Christner (1926) have proposed the name Camp Springs conglomerate for the basal Dockum of this region, but due to the discontinuity inherent in the Dockum sediments it seems best at present to consider the Camp Springs conglomerate as of only local importance.

The skull is of the general phytosaurian form, small in size, longirostrine, and somewhat depressed dorso-ventrally. The extreme anterior position of the external nares suggests a close relationship to the genus *Paleorhinus* Williston. The presence of a lateral quadrate foramen, confirmed by the present specimen, appears now to be characteristic of the genus and offers further evidence of the generic identity of the two specimens. The general proportions of the skull and the unusual construction of the narial structures is similar to that found in *Paleorhinus parvus* Mehl, and it is therefore assumed that the present specimen is more closely allied to that species than to *P. bransoni.* The relative smallness of the present skull (estimated length 550 mm.), the decidedly uncrowded arrangement of the maxillary teeth, and the apparent presence of a persistent parietal foramen attest the immaturity of the individual.

Unfortunately the type specimens of *Paleorhinus* leave much to be desired from the standpoint of preservation. In many respects the present specimen is in better condition than the other two, but because of this fact comparison with them is rendered extremely difficult. Were all three typical *Paleorhinus* skulls equally well preserved many seeming differences in cranial construction might disappear. The writer, however, is of the opinion that such characters as the dorso-ventral flattening of the skull, the exceptionally large palatine foramina, and the moderately elongate posttemporal fenestrae, are of sufficient importance to separate the present form from the previously described species.

ORDER Thecodontia
SUBORDER Phytosauria
FAMILY Phytosauridae
SUBFAMILY Phytosaurininae Camp
Paleorhinus scurriensis n. sp.

HOLOTYPE. Incomplete skull, W. T. Mus. Pal. Col., No. 539.

HORIZON· Basal Dockum, Triassic (Camp Springs conglomerate).

LOCALITY: Two and one-half miles NE of Camp Springs, Scurry County, Texas.

DIAGNOSIS

Small paleorhinid phytosaur; nares anterior, quadrate foramen posterior in position, but visible in lateral aspect. Skull flattened dorso-ventrally. Internarial septum composed of premaxillaries and nasals; septomaxillaries not meeting in the median line. Palatine foramina very large; posttemporal fenestrae elongate. Persistent parietal foramen probably present in immature individuals.

OPENINGS IN THE SKULL

The pronounced anterior position of the narrow and elongate external nares in *Paleorhinus scurriensis* is even more striking than in the previously described species of *Paleorhinus.* They look upward and slightly forward, the posterior borders being somewhat more elevated than the anterior rims. They are transversely wide behind, but become progressively narrow in the anterior direction and finally disappear in narrow slots on the upper surfaces of the septomaxillaries.

The choanae are elongate, narrow openings situated in the palatal vault on either side of a thin but vertically broad median septum. They are situated some distance behind the posterior limits of the external narial openings.

The relatively small antorbital fenestrae are oval in outline. The left opening is slightly larger than the right, the former measuring some 64 mm. by 32 mm. along the extreme axes. They extend forward to a point some 36 mm. behind the posterior limits of the external nares. The openings are only partially visible in lateral aspect although they may be seen in their entirety from above. The lateral areas of the skull immediately surrounding the antero-superior and posterior margins of the fenestrae are sharply depressed in a manner similar to that seen in *P. bransoni.*

The median and posterior rims of the superior temporal fenestrae are preserved on both sides of the skull and indicate that these openings were broadly oval in outline.

The boundaries of the lateral temporal fenestrae are completely lacking, but the extreme thinness of certain portions of the right jugal and squamosal indicate that these areas were not far removed from the borders of the openings on that side of the skull.

A small portion of the frontal rim of the right orbit is preserved. Apparently the orbits were directed upward and outward. The transverse diameter of the frontals between them seems to have been somewhat less than in *P. bransoni.*

Unfortunately the various foramina situated in the cranial region have not been freed from the matrix. Numerous fossae can be seen leading into the general vicinity of them, but their actual positions cannot be determined.

The quadrate foramina, represented by a small portion of the right opening, were visible in lateral aspect although their position was somewhat more posterior than in *P. bransoni.*

The foramen magnum is nearly round in outline.

The posttemporal fenestrae are elongate and vertically narrow. They are larger than in the skull of *P. bransoni*, but are, nevertheless, relatively smaller than in the majority of the larger phytosaur species.

The palatine foramina in *P. scurriensis* are large, elongate openings bounded medially by the palatines and laterally by the ectopterygoids and maxillaries. The total area of each opening appears greater than in any other genus with the possible exception of *Angistorhinus*.

INDIVIDUAL SKULL ELEMENTS

PREMAXILLARIES: Only the posterior portion of the premaxillaries is available for discussion. When seen from above the bones are relatively smooth, strongly rounded transversely on the upper surface, and meet each other in the median line. They are not separated posteriorly by interjection of the septomaxillaries, but, as described in the following paragraphs, continue into the narial opening forming the greater part of the internarial septum. Elsewhere the premaxillaries are excluded from the narial rim by the septomaxillaries and the nasals. They terminate laterally some 40 mm. anterior to the antorbital fenestrae, in a sharp tapering process which separates the nasals and the maxillaries in this region of the skull. Ventrally the premaxillaries are strongly vaulted forming the roof of the anterior portion of the palate. On either side of the median line there is a strong, sharp ridge running longitudinally on the inside of the alveolar trough. Similar alveolar ridges, although reported in the majority of the phytosaurs including *Paleorhinus bransoni*, are, however, described as being "little developed" in *P. parvus* (Mehl, 1928). This ridge dies out posteriorly on the palatal surface of the

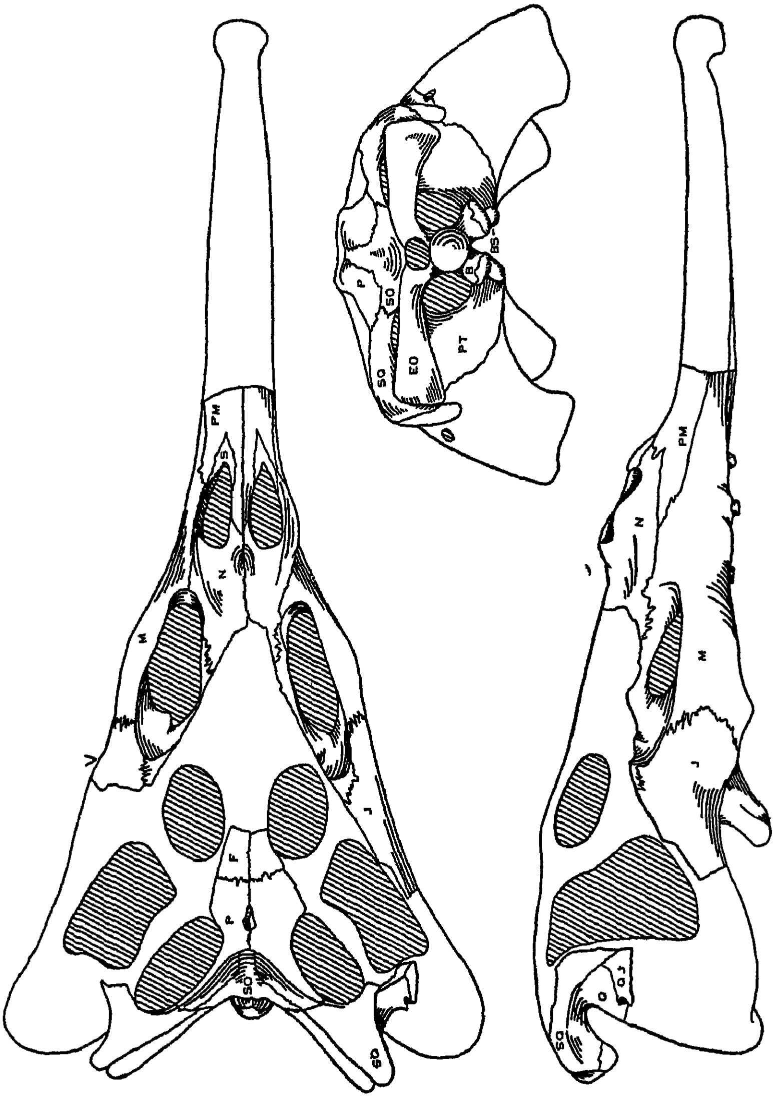

Fig. 1. *Paleorhinus scurriensis*, new species; W. T M. Pal Col., No. 539. Dorsal, lateral, and posterior aspects of skull. Approximately ¼ nat. size. Abbreviations: B, basioccipital; BS, basisphenoid; EC, ectopterygoid; EO, exoccipital; F, frontal; J, jugal; M, maxillary; N, nasal; P. parietal; PA palatine; PM, premaxillary; PT, pterygoid; Q, quadrate; QJ, quadratojugal; S, septomaxillary; SO, supraoccipital; SQ, squamosal. V indicates area through which section (Figure 3) was taken.

maxillary. Only one possible premaxillary alveolus is preserved in each bone. Posteriorly the premaxillary-maxillary sutures converge on the median line causing the premaxillaries to become narrow in this direction. They project into the choanae for a distance of 10 mm. at which point they have been destroyed by a transverse fracture through the skull. There is no contact with the palatines on the palatal surface.

As in *P. parvus* the relatively thick internarial septum rises somewhat above the level of the elevated narial rim. It is composed principally of the postero-mesial extensions of the premaxillaries which are overlapped laterally by the thin mesial wings of the septomaxillaries. The nasals project between the nares for a distance approximately one-fourth of the length of the septum.

In the majority of the phytosaurs this septum is formed by the septomaxillaries and the nasals. The premaxillaries are usually excluded from it by the juncture of the septomaxillaries in the median line. The preservation of the present specimen is such that when two adjacent bones have been damaged by weathering, a thin lamina of concretionary matrix remains standing to mark the position of the suture. These concretionary partitions are clearly seen in the region of the nares, and there can be little doubt that the condition exists as figured.

In his restoration of the skull, Lees (1907) indicates that the internarial septum in *P. bransoni* is composed altogether of the nasals which extend well forward anterior to the external nares and bound those openings completely. No septomaxillary elements are shown in the figures. Mehl (1928), however, has demonstrated the presence of these bones in the skull of *P. parvus* and their position in *P. scurriensis* cannot be questioned. Insofar as can be determined the relationship of the septomaxillaries to the premaxillaries and the nasals and the general construction of the internarial septum is similar if not identical in the two latter species, and there is, therefore, strong reason to expect similar conditions in *P. bransoni.*

SEPTOMAXILLARIES: The septomaxillaries are relatively small V-shaped elements that form the anterior rims of the external nares. They project forward between the posterior wings of the premaxillaries for a distance of only 18 mm. The posteriorly directed mesial process of each bone extends well into the naris where it laps against the septal portion of the premaxillary. Insofar as the writer is aware, a similar relationship is to be seen elsewhere only in *Paleorhinus parvus.* Mehl (1928), however, was unable to determine the posterior extent of the septomaxillaries in *P. parvus,* and therefore it is not known whether these elements joined each other in the median line in that form. In the present specimen, however, it is certain that they do not.

MAXILLARIES: The maxillaries are rather massive and elongate. Their maximum anterior extension on the lateral surfaces of the skull cannot be determined. In palatal aspect they form most of the anterior borders and part of the lateral margins of the choanae. They do not meet in the median line of the palate, being everywhere separated by the posterior continuation of the premaxillaries. Laterally the maxillaries occupy large areas on the sides of the skull and contribute greatly to the borders of the antorbital fenestrae. In fact, from the preservation of the specimen, it would appear that the maxillaries completely surround these openings. It seems probable, however, that at least a small part of the upper margins of the fenestrae would be contributed by the lacrimals. Beneath this opening the maxillaries are vertically thickened and become somewhat convex downward. The maxillaries contain 16-17 rounded alveoli of which the anterior five or six are decidedly the smallest. They are separated by transverse partitions and it appears that the teeth were not greatly crowded, a feature which has been interpreted as a youthful character in *Machaeroprosopus* (Colbert, 1947).

JUGALS: The jugals are poorly preserved and can be only partially described. They articulate anteriorly with the maxillaries by means of angular interdigitating sutures. These sutures run upward and backward along the ridges created by the depression of the lateral skull surfaces just posterior to the antorbital fenestrae. In lateral aspect the ventral outline of the jugals is strongly arched permitting a wide view of the postero-ventral processes of the ectopterygoids. The relationships of the jugals to the orbits and lateral temporal openings cannot be accurately determined.

No recognizable portion of either lacrimal is preserved.

NASALS: The nasals are massive bones surrounding the external nares laterally and posteriorly. They are rugosely sculptured dorsally, but are relatively smooth on the lateral surfaces. They project slightly between the narial openings, forming not more than one-fourth of the internal septum. Posterior to the nares the nasals are depressed and transversally constricted. Below the narial rim there is a moderate depression of the lateral skull surfaces which together with the constriction of the dorsal surface posterior to the nares lends a rather bulbous appearance to the nasal prominence. In lateral aspect the bones become wedge-shaped anteriorly where they separate the premaxillaries from the septomaxillaries for a short distance. They terminate laterally slightly in advance of the external nares. The nasals do not contribute to the borders of the antorbital fenestrae, but are separated from those openings by thin, backwardly directed processes of the maxillaries. Due to the loss of the dorsal region of the skull posterior to the nasal prominance the relationships of the nasals to the other bones of the skull cannot be determined.

FRONTALS The frontals are poorly preserved and little can be said of them. The transverse diameter between the orbits appear to have been slightly less than in *P. bransoni*, and as a result the orbits probably looked more directly upward than in that form. The frontals meet the parietals along an almost straight transverse interdigitating suture.

PARIETALS: The parietals are proportionately of about the same size as in the type of *P. bransoni* although the maximum transverse diameter appears somewhat less than in that form. The sutural contacts are largely obscured in the present specimen, but the median and frontal sutures can be delimited to some extent. When seen from above the parietals appear as postero-median elements of roughly Y-shaped outline, the upper surfaces of which are rugosely sculptured. They are deeply excavated on either side by the superior temporal fenestrae. Posteriorly the parietals separate, each bone sending backward a wing-like process which lies above the supraoccipital and meets the antero-mesial branch of the squamosal. As a result of destruction of the upper surface of the posttemporal arcade on either side of the skull, the exact point at which this latter juncture occurs cannot be determined. It is the opinion of the writer that the suture must lie at about mid-length of the posttemporal bar. It is certain that it does not occur mesial to that point.

Some 12 mm. from the fronto-parietal suture an abnormally deep pit of roughly ovate outline excavates the parietals in the median line. It is certain that this pit, although much deeper than the adjacent sculpture, does not extend into the cranial cavity. Its relations to the internal structure

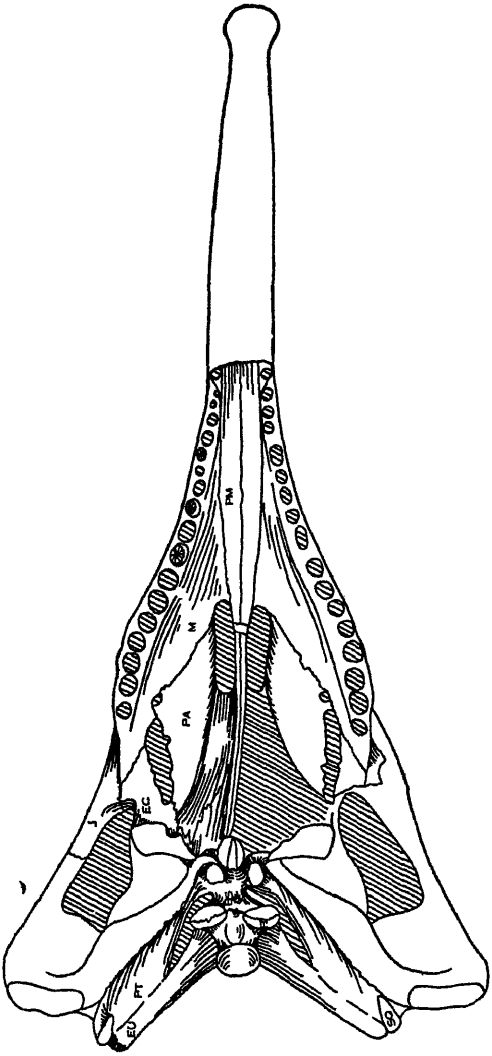

Fig. 2. *Paleorhinus scurriensis*, new species; W. T. M. Pal. Col, No. 539. Palatal aspect of skull. Approximately 1/4 nat. size. For abbreviations see Figure 1.

of the cranium cannot be observed. The writer, therefore, cannot state definitely that this pit represents a pineal foramen, but its position is strongly suggestive. A pineal foramen is reported in *Mesorhinus* and Camp (1930) has demonstrated its presence in a juvenile specimen of *Machaeroprosopus lithodendrorum.* In view of this evidence and the fact that the epiphysis is strongly developed in adult individuals of *Machaeroprosopus* (Camp, 1930), vestiges of the pineal opening might be expected in immature members of the Phytosauria.

SQUAMOSALS: The squamosals are poorly preserved. It can be seen, however, that they are overlapped posteriorly by the paraoccipital processes of the exoccipitals. They articulate laterally in diagonal sutures with the quadrates. Apparently they form about one-third of the postero-mesial borders of the superior temporal fenestrae, but as pointed out above, the exact position of the parieto-squamosal sutures cannot be determined. The squamosals and quadrates are deeply excavated in the vicinity of the otic foramen. Lees (1907) described this area as containing an otic foramen in *P. bransoni.* In the present specimen, however, this area is well preserved and there is no indication of any structure that might have bounded the opening on the ventral side. It is, therefore, concluded that while the otic opening is deeply excavated in *P. scurriensis* it was present as a deep notch rather than as a completely bounded foramen, and the writer is inclined to agree with Mehl (1928) and others in questioning the accuracy of Lees' interpretation of this region in the genotype.

QUADRATES AND QUADRATOJUGALS: Only a small portion of the quadrate and quadratojugal from the right side of the skull is available for discussion. Strong indications of a quadrate foramen as shown by Lees in *P. bransoni* are to be seen on the postero-lateral surface of the quadrate. The opening is clearly visible in lateral aspect, but it is situated somewhat nearer the posterior edge of the quadrate than in that species.

PALATINES AND PTERYGOIDS. The palatal vault with its extremely thin and delicate bones is by no means well preserved in the present specimen. In fact, it has been deemed inadvisable to attempt the removal of the matrix from this region on the left side of the skull. A longitudinal fracture has, however, caused the exposure of the sides and roof of the palatal vault on the right side and, although badly broken, the palatine and pterygoid can thus be delimited with some degree of certainty. The posterior limits of the choanae cannot be determined in the specimen.

The palatines and pterygoids form the greater portion of the high steeply vaulted palate posterior to the choanae. The former elements are of the characteristic phytosaurian construction with the usual horizontal and vertical plates set off from each other at angles approaching 90 degrees. Laterally and anteriorly the palatines are in sutural contact with the maxillaries; postero-superiorly they are overlapped by the pterygoids, whereas postero-laterally they meet the ectopterygoids. When seen from below, the hoizontal plate of each pterygoid is broadly triangular in outline. The base of this triangle is somewhat rounded and thickened whereas the posterior leg forms the internal boundary of the pataline foramen, which in *P. scurriensis* is slightly less than one-half (60 mm.) of the total length of the palatine. Anteriorly the mesial border of the horizontal plate forms the postero-lateral rim of the choana, but other relationships of the vertical plate in this region cannot be determined. Each palatine terminates anteriorly, on the lingual surface, in a sharp pointed process which wedges out in the maxillary at a point just behind the anterior borders of the choanae.

The badly broken right pterygoid only is preserved in the palatal region. It forms part of the vertical sides of the palate and enters into the palatal roof to an undetermined extent. Along the median line of the palate there is a rather sharp, ventrally directed carina which is formed, at least in part, by the joining of the downward turned mesial edges of the pterygoids. This carina continues forward into the choanae and there probably meets the septal portion of the premaxillaries although the imperfection of the specimen in this region makes it impossible to state the exact position of the juncture. When seen from behind both pterygoids appear as broad-wing-like plates of bone. There is no indication of the presence of the quadrate suture on the preserved surfaces of either bone. Below, downwardly and anteriorly directed flange-like processes fold around the anterior and lateral surfaces of the basipterygoid processes.

There is no evidence of vomerine sutures and the writer is inclined to believe that these elements did not appear in the palatal aspect in *P. scurriensis.*

ECTOPTERYGOIDS: The right ectopterygoid is partially preserved. The antero-lateral process articulates with the maxillary and jugal and surrounds the postero-lateral half of the palatine vacuity. In lateral aspect the bone projects downward well below the lower border of the upwardly arched jugal. It underlies the outer process of the pterygoid.

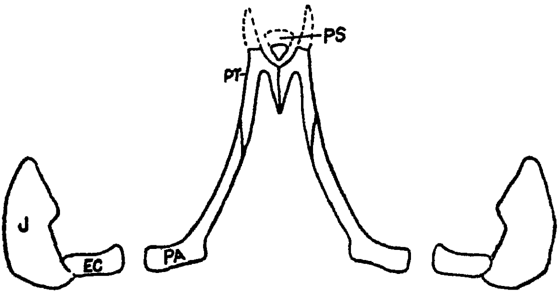

Fig 3 *Paleorhinus scurriensis,* new species; W. T. M. Pal. Col, No. 539. Transverse section through skull (V, Figure 1), somewhat diagrammatic. PS, parasphenoid process. ½ nat size.

SUPRAOCCIPITAL: The supraoccipital is a relatively large triangular element situated in the postero-mesial notch of the skull. From above it is more completely visible than in *Paleorhinus bransoni,* approaching in this respect, the condition found in *Promystriosuchus ehlersi.* It forms only a small part of the upper boundary of the foramen magnum and extends upward terminating against the posterior eave-like extension of the parietals just below the level of the skull roof. Laterally the bone extends outward on either side in narrowing wing-like processes to articulate with the squamosals on the posterior face of the skull about mid-length of the post-temporal arcade. Supero-laterally it is bounded by the parietals whereas ventrally it unites with the exoccipitals along horizontal sutures.

EXOCCIPITALS: The exoccipitals are of the characteristic phytosaurian form although the paroccipital processes appear to be somewhat more elongate than in the majority of the genera. Except for the small mesial portion, contributed by the supraoccipital, they bound the foramen magnum completely and extend downward onto the occipital condyle forming a major part of the neural surface of that process. The paroccipital processes are relatively straight, extending strongly backward and outward to terminate in rather broadly expanded buttresses against the squamosals. The elongate but vertically narrow posttemporal fenestrae are situated at about mid-length of this process.

BASIOCCIPITAL: The basioccipital comprises the greater part of the rounded knob-like occipital condyle which is set off from the remainder of the bone by a strong but relatively short neck. It is excluded from the floor of the foramen magnum by the junction of the exoccipitals in the median line. Ventrally the basioccipital is divided along the median line by a deep groove thereby forming the posterior halves of the strongly developed roughened tubera. Supero-laterally it is in contact with the exoccipitals whereas antero-ventrally it joins the basisphenoid.

BASISPHENOID: The basisphenoid is a large, irregular median element situated just anterior to the basioccipital with which it is cojoined along a very tenuous suture. The connection between the two bones is much less stout than in the larger phytosaurs. Postero-ventrally this suture runs transversely across the large downwardly and anteriorly projecting tubera of which the basisphenoid forms the anterior one-half. Antero-ventrally the bone is again expanded on either side of a deep median pit into two strong, downwardly and laterally projecting basisphenoid processes which are enclosed anteriorly by the median, recurved flanges of the pterygoids. Much of the matrix still remains in this region of the skull and other features of the basisphenoid and adjacent elements are therefore obscured from view.

LATEROSPHENOIDS: Insofar as they can be seen the laterosphenoids underlie the frontals and parietals medially. They are broken on either side of the skull, but it is apparent that they also underlie the postfrontals to some extent. Posteriorly they meet the prootics, but are excluded from contact with the supraoccipital by the junction of prootics and the parietals in front of those bones. The bones are rather bulbous in appearance beneath the parietals, but taper gently inward anteriorly beneath the frontals.

PROOTICS· The prootics are inadequately exposed, but it can be seen that they are relatively large. They are bounded above by the supraoccipital and antero-posteriorly by the parietals.

GROWTH FACTORS IN THE GENUS *Paleorhinus*

The size and robustness of the type skull of *Paleorhinus bransoni* indicate that the individual was adult or nearly so whereas there can be little doubt as to the immaturity of the present specimen. The question of growth factors in the genus *Paleorhinus* may therefore be examined.

Colbert (1947) has attributed certain variable characters in *Machaeroprosopus* to factors of growth. The extent to which these criteria (robustness, prenarial crest, posterior expansion of premaxillaries, spacing of teeth, and proportion of prenarial length to skull length) can be applied to the genus *Paleorhinus* is not completely clear, but most of them appear to be of at least minor importance.

Differences in robustness are clearly evident in skulls of *P. bransoni* and *P. scurriensis*; the latter is more lightly constructed in every detail. The presence of prenarial crests within the Phytosaurininae appears to be peculiar to *Machaeroprosopus*, *Phytosaurus*, and in somewhat less degree to *Leptosuchus*. The rostrum in *Paleorhinus* apparently underwent some increase in overall massiveness during growth, but nothing resembling the crests of the above named genera is evident. A posterior expansion of the premaxillaries is present in *P. bransoni*, but is of less magnitude than in *Machaeroprosopus*. This region is missing in the present specimen. The teeth in *P. bransoni*, while not approaching the crowded condition found in the adult *Machaeroprosopus* skull are, nevertheless, more closely spaced in the Texas specimen. The changes in proportion of the prenarial length to skull length cannot be established on the basis of available material. The presence of a parietal foramen has been discussed above.

Two features observed in the study of the skull of *P. scurriensis* are perhaps significant although their variability is perhaps most important from the phylogenetic standpoint The condition of the septomaxillary as described in the foregoing discussion is not seen in the more advanced forms. Two small bones in the same relative positions as in *P. scurriensis* are figured in *Mesorhinus* (Jaekel, 1910) and there is reason to believe that these elements represent the septomaxillaries in a much reduced and primitive condition. If *Mesorhinus* is accepted as a primitive structural type, and *Paleorhinus* is considered as an intermediate stage, the proportional changes occurring in the phytosaur skull during the Triassic will clearly demonstrate the development of the remarkably large median septal elements of such advanced American forms as *Machaeroprosopus.*

The second feature, the large size of the palatine foramen, is a feature which appears to be peculiar to the more primitive Phytosauria (with the possible exception of *Promystriosuchus*). It is therefore assumed that immature skulls in which a parietal foramen still persists might still reflect this primitive condition of the palatine foramen. It is certain that the openings are larger in *P. scurriensis* than in *P. bransoni.*

The effect of sexual dimorphism upon the variability of these characters is at present unknown. They may be summarized as follows:

CHARACTER	YOUTHFUL CONDITION	ADULT CONDITION
Robustness	general trend toward	increase in massiveness
Prenarial crest	absent	absent
Posterior expansion of Pmx	not observed	present, but of small magnitude
Spacing of teeth	wide	narrow
Prenarial length/skull length	undetermined	
Size of septomaxillaries	small	larger?
Palatine foramen	large	smaller
Pineal foramen	persistent	obliterated

COMPARISONS WITH OTHER FORMS

A comparison of the skull of *Paleorhinus scurriensis* with the two previously described *Paleorhinus* species is tabulated below:

CHARACTER[3]	*P. bransoni*	*P parvus*	*P scurriensis*
Nares anterior	3	2	1
Squamosal short, heavy	1	—	2
Quadrate foramen visible in lateral view	1	—	2
Otic opening a notch	1	—	2
Skull flattened dorso-ventrally	2	—	1
Internarial septum composed of Pmx. and Na	—	0	1
Septomaxillaries not meeting in median line	—	0	1
Palatine foramina large	2	—	1
Posttemporal fenestra elongate	2	—	1
Number of maxillary teeth	16	17	16–17

[3] 1-3 indicates degree of development, No. 1 indicates maximum development 0 indicates a probable difference, but degree not completely observed.

Differences between the present specimen and skulls of *Leptosuchus*, *Brachysuchus*, and *Machaeroprosopus* are to be found most abundantly in the above table. In fact, no points of complete agreement with the listed characters are to be found except in the open otic notch. *Mystriosuchus* and *Rutiodon* are completely different in appearance and general proportions. *Angistorhinus* is a relatively primitive genus that exhibits many similarities to the present skull. Most of these may be considered as of only superficial importance and the characters of the subfamilies are adequate for definitive purposes.

There is much similarity between the skull of *Paleorhinus scurriensis* and that of the nearly contemporaneous Texas form *Promystriosuchus ehlersi* Case. Both are of relatively small size with slender rostrae. The nares are anterior in each, but they are somewhat more elevated in the latter. The depression of the facial and posterior cranial portions is more pronounced in *Promystriosuchus.* The posttemporal arcade is apparently in about the same stage of development in both skulls although the squamosals are more strongly developed posteriorly in *Paleorhinus scurriensis.* The number of teeth is approximately the same. The elongate and broadly exposed parasphenoid process of *Promystriosuchus ehlersi* is remarkably different from that in all other phytosaurs, including *Paleorhinus.* The palate is more steeply vaulted and the palatine foramina are much larger in *Paleorhinus scurriensis.*

SYSTEMATIC POSITION

The present specimen is in complete agreement with the characters listed by Camp (1930) as typical of the subfamily Phytosaurinminae and is obviously primitive within that group. Its similarity to the paleorhinid skull as exemplified in *P. bransoni* is obvious, but its position within the genus is obscured by the fact that it pertains to an immature individual in which important comparative features are not fully developed. For this reason it seems advisable to defer further decisions on the systematic position of *P. scurriensis* in the hope that future discoveries in the Dockum group may provide an adult individual.

GEOLOGICAL IMPORTANCE

Recognition of the genus *Paleorhinus* in Texas again raises the question of terrestrial Triassic correlation in western North America. Detailed correlation, however, is hazardous even within the relatively small areas of continuous Dockum exposures. In many instances it is impossible to prove the age equivalence of exposures that are separated by only short distances. It is not surprising, therefore, that attempts to integrate the present stratigraphic information concerning the Dockum and other Triassic deposits have met with little success.

In view of the paucity of identifiable invertebrate fossils in the continental Triassic redbeds and the extreme degree of lithologic facies variation within the sediments, the evidence afforded by vertebrate remains becomes of great stratigraphic importance. Numerous discoveries of these remains have provided surprisingly few instances of genera common to two or more areas. Several supposed occurrences of faunal duplication have been reported, but in most cases the identifications have been based upon inadequate material. The validity of such occurrences must, therefore, be accepted with scepticism.

Two phytosaur genera, *Angistorhinus* and *Paleorhinus*, originally described from the Popo Agie[4] of Wyoming, are known to be present in the Dockum of Texas; the former through at least two occurrences in the Big Spring area (Stovall and Whorton, 1936; Gregory, 1944) and the latter through the present specimen. Neither of these forms has yet been reported from Dockum deposits farther to the north which are presumably somewhat higher stratigraphically than the exposures in Scurry and Howard counties. In fact, the skull of *Paleorhinus scurriensis*, coming from a level just above what is considered as the Permian-Triassic boundary in Scurry County, may represent the oldest phytosaur yet reported from the Dockum Exact information regarding the stratigraphic position of the Texas *Angistorhinus* specimens is not available, but they must have come from a level somewhat higher than that which yielded the present specimen. Information concerning the relative positions of the two genera in the Popo Agie is likewise lacking, but insofar as the writer can determine, the two forms were never found in association.

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[4] Reeside (1929) proposed that the name Jelm formation be substituted for the Popo Agie formation, basing his argument upon structural, stratigraphic, and lithologic considerations Others later concurred in this opinion. Without regard to the apparent desirability of such a substitution, however, it must be stated that the term Popo Agie formation has found wide usage as a vertebrate horizon within a section of beds relatively barren of identifiable vertebrate remains and as such still possesses considerable value for purposes of vertebrate paleontologic correlation. The term as here employed refers only to that portion of the Jelm formation from which diagnostic vertebrate remains have been described.

In general, proposed correlations betwen the Dockum and Popo Agie have been couched in rather general terms, but the trend has been to indicate an *in toto* age equivalence for the two. The writer is of the opinion that the equivalence of the Popo Agie and the lower part of the Dockum group can now be demonstrated within acceptable limits of error. However, this equivalence which is most accurate in the Howard and Scurry County area must become less distinct laterally. The vertical limits of equivalence cannot be accurately determined, although it is apparent that the Popo Agie sediments represent a much shorter period of deposition than do those of the Dockum group.

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REVIEWS

Cybernetics; by Norbert Wiener. Pp. 194. New York, 1948. (John Wiley & Sons, $3 00) —The title of the book, we are told by its author, means "control and communication in the animal and the machine"; this happy coinage of a new word for a new subject should be applauded by less literate scientists as scoring against those who have scornfully concluded that scientists need lessons in linguistics.

Wiener's book makes significant contributions to several fields. In physics it calls attention to a change in the concept of time which attended the passage from Newtonian to statistical mechanics. The author holds that Newtonian time is reversible, a kind of time in which nothing new can ever happen. Statistical mechanics, particularly in the form of quantum theory, has significance only for an irreversible time ("Gibbsian time" according to Wiener—a rather strange baptism), a time "in which there is always something new." This irreversible time is essentially the one advocated by Bergson as characteristic of biological phenomena, and the point is made that physics, having achieved a stage in which it can build automata that resemble organisms, must take seriously the irreversible time of Bergson.

Nowhere is the interesting connection between Gibbsian physics and Lebesguian mathematics so clearly outlined as in the present book. And the statement bristles with challenges unpleasing to the devotee of Gibbs. Thus, we find an appreciative acknowledgement of Gibbs' discovery of a relation between time averages and space averages, followed by a reminder that "in the method by which he tried to show this relation he was utterly and hopelessly wrong." It was a surprise to the reviewer that Gibbs ever tried to establish such a relation; for it appears from his writings that he contented himself with the brilliant postulation of what he called the canonical ensemble, and that he did not worry about ergodic problems. At any rate, Wiener's second chapter brings the logical and mathematical situation surrounding the ergodic hypothesis soundly up to date.

A review of the author's important theories of statistical prediction and of filters is included in the book. This will be of some interest to mathematicians and theoretical physicists, and will have but little appeal to most readers. The value, even to the specialist, of the mathematical portions is reduced by many unfortunate typographical errors which make the detailed examination of analytical steps rather exasperating in several important sections of the book.

The engineer will gain a deepened insight into the workings of feedback mechanisms. There is in chapter 4 a beautiful discussion and a simple classification of the various types of feedback systems from a general and formal point of view. A great deal of information about the action and construction of calculating machines is offered, and interesting comments on numerous problems of communication engineering are found throughout

The book attempts to appraise neuro-physiology in a new way. Adopting a forthright behavioristic attitude, Wiener "explains" the workings of the brain in terms appropriate to the action of modern calculating machines. There is apparently an extensive parallelism between neural anomalies, or even psychopathic conditions, and the malfunctioning of electric calculators; this parallelism is skillfully exploited and made highly informative and suggestive.

Considerable space is given to the social consequences of our ability to construct automata. Enormous popular interest has seized and will continue to seize upon this lurid aspect of cybernetics, but there is no need to dwell upon it in a brief review.

As to the bearing of his researches upon philosophy, the author makes two basic claims. The first is in his thesis that all regularities in nature, all laws, are essentially of a statistical sort. Exact laws, like that of gravitation, are limiting cases, are in fact not exact in detail, and represent matters of little contemporary interest, to judge from the tenor of Wiener's argument. Confirmation for this thesis is seen in the quantum theory of the atom In my opinion such argumentation is rather artificial, suggesting an extreme and self-defeating form of empiricism. For while it is true that the interpretation of quantum laws is a statistical one, the laws themselves, like Schrödinger's equation, are still exact. Even if one admits, as one must, that detailed observations scatter within a certain range and are in accord with the laws of chance, important parts of our experience nevertheless remain mysteries. For there are still laws of nature, and presumably exact laws, which regulate phenomena "in the mean." Thus the attitude of *nil mirari,* displayed throughout the book with respect to matters of traditional philosophy, comes to grief at this point; for it is just as hard to account for the statistical laws of quantum physics as it is to account for the laws of classical physics. The shift of emphasis effected by Wiener's thesis, while timely and useful, seems to solve no philosophical problem.

The second point of importance is Wiener's insistence that a proper understanding of the meaning of information, defined as negative entropy, provides a uniform basis for thermodynamics, communication engineering, and psychology. Here again it is typical that no place is given to consciousness, which the author regards as superfluous. In this, many scientists will follow him. But there seems to be a strange defect in the whole philosophy of cybernetics which arises from its concern over behavior in a very narrow sense. Suppose we admit that information can be formally defined in terms of the varying amplitude of some carrier. Does it not matter greatly what this carrier is? Every science is intensely interested in the substratum which communicates its signals; it is important indeed whether they are acoustic, seismic, electromagnetic, or perchance gravitational. The philosopher will want to ask Dr. Wiener to commit himself more clearly than he does in this book on the *nature* of the information, which he formally defines, that roams around in what is called the human mind Until he does he teaches an interesting analogism quite similar to one well-known in physics which embraces mechanical, acoustic and electrical vibrations.

Perhaps this is all he intends, and it would be a noteworthy scientific achievement, amounting in the end to something like the following. We now have a set of mechanical, acoustic and electrical analogies (mass, stiffness of a spring, force, etc). By means of them, the solution of a problem in mechanical vibrations becomes at once a solution of a problem in acoustics and a solution of another problem in alternating current networks. This circumstance is most fortunate and scientifically useful. But it does not allow us to say that mechanics, acoustics and electricity are basically the same thing Wiener's contribution is, it seems, in the suggestion that the known analogies can possibly be extended to certain phenomena in the animate world. To drive the philosophical implications of cybernetics beyond that point is at present unwarranted.

HENRY MARGENAU.

Practical Spectroscopy; by GEORGE R. HARRISON, RICHARD C. LORD and JOHN R. LOOFBOUROW Pp. xiv, 605. New York, 1948 (Prentice-Hall, $6.65).—The first practical spectroscope was developed by G. R. Kirchhoff and R Bunsen in 1859 They showed that the spectroscope could be used for chemical analysis, and with their instrument they discovered several new elements and demonstrated the existence of many of the then known elements in the sun. From this beginning spectroscopy has had a tremendous growth, especially in the last thirty years during which guiding quantum theories have been available. Today the spectrograph serves as a powerful research tool in all branches of science.

At the Massachusetts Institute of Technology the authors, who are a physicist, a chemist, and a biophysicist respectively, have been operating what is probably the largest and best equipped spectroscopic laboratory in this country. In this book they synthesize their viewpoints so well that all who use spectroscopic techniques, no matter what their main scientific interest, will find in it much of value. No topics are treated in an advanced way, but references are always given so that the reader may without delay delve further into the literature of any subject. The authors avoid the historical approach. They give a comprehensive view of the present status of all the instruments and techniques of ordinary spectroscopy, together with the main results of the supporting theory in every case. They do not discuss the recently very active field of microwave spectroscopy.

After a first chapter which gives a birds'-eye view of spectroscopy as a scientific tool, there follow chapters on the various types of spectrographs and their adjustment, care, and illumination, photography of the spectrum, light sources of all kinds, and the identification of spectrum lines. The next two chapters summarizing the quantum mechanical interpretation of atomic and molecular spectra and the correlation with atomic and molecular structure form a clearly written introduction to these modern revolutionary developments. Further chapters on the measurement of spectral intensities, photographic photometry, absorption spectrophotometry, both qualitative and quantitative spectrochemical analysis, spectroscopy of the infrared and ultraviolet, Raman spectroscopy, and the use of the various types of interferometers are packed full of useful information.

Appendices giving the sensitive lines of each of the elements arranged both according to elements and to wave-lengths should prove of considerable value because of the compact arrangement and the inclusion of recent measurements. Numerous photographs and line drawings form an attractive feature of the book. The style is direct, forceful and economical.

The authors of *Practical Spectroscopy* are to be congratulated for their production of such a complete, well-written account of modern experimental spectroscopy. Everyone concerned with any aspect of the subject will find this book indispensable.

W. W. WATSON

The Interpretation of Spectra; by WM M. VENABLE Pp. 208; 12 figs, 73 tables New York, 1948 (Reinhold Publishing Corp., $6.00).—This small book will delight the iconoclast. It is an assiduous attempt to account for *all* optical and x-ray spectra on classical electromagnetic theory. Quantum theory and quantum electrodynamics are dismissed as being hypotheses having little relation to observed data. The problem which first precipitates this drastic philosophy is the correlations between the principal spectrum of hydrogen and its many-line spectrum. By treating the hydrogen atom as a combination of three "sub-atoms" the author claims to have accounted completely for the lines in the hydrogen spectrum; the "energy levels" (a term always to be employed in quotation marks) that result from this model are relatively few in number but can intercombine in rather arbitrary ways to account for the spectral lines. This theory is then extended to helium and the more complex atoms, all of whose nuclei are made up of various combinations of "sub-atoms" in a way that is amazing, to put it mildly. (Incidentally, it is the contention of the author that the alpha particle is *not* a helium nucleus, Rutherford's work notwithstanding.) Inevitably, the proposed model of the atom must meet up with the vast mass of data derived from nuclear reactions, but it appears sufficiently flexible to hurdle that barrier, even if somewhat arbitrarily.

It should be noted that the author is by no means unaware of contemporary theory and experiment; he simply does not subscribe to such work. He derives his philosophy from a remark which he ascribes to Max Planck:

> "Since nothing probably is a greater drawback to the successful development of a new hypothesis than overstepping its boundaries, I have always stood for making as close connection between the hypothesis of quanta and classified dynamics as possible, and for not stepping outside of the boundaries of the latter until the experimental facts leave no other course open."

Admittedly, the gymnastics required to remain with the confines of classical theory are strenuous (the book contains sixty pages of tables), but the author has mastered them to his own satisfaction, if not to this reader's. RICHARD F. HUMPHREYS

Organic Reactions. Volume IV. Edited by ROGER ADAMS, Editor in Chief, WERNER E. BACHMAN, A. H. BLATT, LOUIS F FIESER, and JOHN R. JOHNSON. Pp. viii, 428. New York, 1948 (John Wiley and Sons, $6.00).—This volume, the fourth to appear in a series devoted to the application of specific organic reactions, is entirely worthy of its precursors. Since the appearance of the first volume in 1942, *Organic Reactions* has come to be regarded as indispensable to the organic chemist and each new volume is eagerly anticipated. The present volume consists of eight chapters, the first two of which are devoted to the Diels-Alder reaction and are written by Milton C. Kloetzel and H. L. Holmes respectively. Other chapters are: The Preparation of Amines by Reductive Alkylation, William S. Emerson; The Acyloins, S. M. McElwain; The Synthesis of Benzoins, Walter S. Ide and Johannes S Buck; Synthesis of Benzoquinones, by Oxidation, James Cason; The Rosenmund

Reduction, Erich Mossettig and Ralph Mozingo; and the Wolff-Kishner Reduction. Following the usual form established in the first volume, each chapter considers the mechanism, the scope, and the experimental details of the reaction Several illustrative preparations are given in detail and a complete tabular survey of the applications of the reaction reported in the literature supplements each chapter. The critical evaluations of the reaction details by the authors are particularly valuable, coming as they do from wide first hand experience.

Perhaps the only criticism which this reviewer is inclined to make of this highly commendable series, is that certain of the chapters to appear are of less general significance than could be desired; moreover, in the case of several chapters, e.g., the Diels-Alder reaction, comprehensive treatments of the topics have appeared elsewhere. The reviewer feels that ideas for future chapter topics might be welcomed by the editors and suggests the Hoesch and Gatterman syntheses of aromatic ketones and aldehydes, or perhaps the degradation of amines by the methods of Hofmann, von Braun and Emde. ALFRED H FRYE

The Chemical Constitution of Natural Fats; by T. P. HILDITCH. Pp 554. New York, 1947, John Wiley and Sons, Inc. ($9.00). The author has classified the natural fats on the basis of their sources, and has emphasized the close relation between component acids in fats and their origins. Readers can thus get many new hints towards deciding the direction of their later work in this field. In this second edition, this principle of classification is also adopted and further developed.

The natural fats are dealt with separately in three groups: those of aquatic flora and fauna, land animals, and vegetables. These are further classified under species and family, as well as under different parts of the body in some instances. On the other hand, about one hundred important individual fatty acids are discussed in another chapter, thus giving a clearer understanding of the "building-stones" of fats. The natural forms of glycerides in vegetable and animal fats are discussed in three chapters, including the crystallization method for their separation, although at present complete isolation is still not successful. Recent data of synthetic glycerides are also introduced for convenience of comparison with the natural ones. Many experimental techniques, data and apparatus, such as fractional distillation in vacuo, separation of acids by the lead salt, as well as by the newly developed lithium salt method, and low temperature crystallization, are described in detail. Moreover, recent important literature published after 1940 are collected up to 1945 (although some data published in the Orient were not available at that time), thus giving fresh information to researchers in this field Because of the form of this book, somewhat complicated indexes are needed, so that in this edition an index of plant families was added. The old indexes, such as the index of fats and waxes of different origins, index of individual fatty acids, index of individual glycerides have been retained

The richness of information and accurate data of this book will make it easier for readers to do better future work. This is a good reference for the oil and fat chemists as well as botanists, zoologists and bacteriologists. However, in my opinion, it is desirable and might be more stimulating if there were present a chapter dealing with those sterols obtained from natural fats, although they are not, as the author pointed out already, a component of the fat — the glycerides of fatty acids. The interest which the earth's scientists are beginning to take in the organic components of muds and sediments may recommend this reference book to their attention.

S. L. LIU

When the Earth Quakes; by JAMES B. MACELWANE; S.J. Pp. xi, 228; 252 figs Milwaukee, 1947 (The Bruce Publishing Co. $5 00) —In this book one of the foremost seismologists attempts to explain his science to educated laymen The general plan of the treatment resembles that followed by several other books in the same field. Starting with a discussion of earthquake behavior, documented by the recorded effects of 18 major shocks in modern times, the author proceeds to an explanation of causes; of methods and techniques used in the scientific study; of the earth's internal structure as deduced from earthquake records, and of practical applications of seismic principles in the search for mineral deposits Considerable attention is given to the engineer's interest in seismology, in connection with problems of earthquake-resistant construction. A chapter is devoted to rock bursts in underground workings, and another to microseisms caused by storms at sea. An appendix of 20 pages outlines some of the principles of physical geology.

Without question any intelligent reader can get from the book much interesting and accurate information. The style is clear, essentials of the subject are well covered, and the language is reasonably non-technical. An extensive glossary defines terms with which most lay headers are not familiar. Numerous good illustrations supplement the text, but some of the figures are open to criticism. Of more than 40 maps, only two are equipped with graphic scale, and six others have geographic coordinates; thus most of the maps lose much of their value to readers Many halftones showing seismographs and other technical instruments are of dubious value, since in general these pictures give no hint of principles on which operation of the instruments depends. A few well constructed diagrams would be far more serviceable.

Scientists other than those well versed in seismology will find that the book has much instructive reading, and that it will give service as a reference work for some subject matter. One of the appendices gives a long list of publications dealing with earthquakes and seismic studies. Unfortunately these are not arranged in alphabetical order of authors. CHESTER R. LONGWELL

Fundamentals of Physical Science; by KONRAD B. KRAUSKOPF. Pp. xii, 676; 330 figs. Second edition, New York, 1948 (McGraw-Hill Book Co., $4.50).—Krauskopf's textbook attempts two things: first to survey all the physical sciences—astronomy, physics, chemistry, geology—presenting in brief form the principal conclusions of each that the intelligent layman would profit from knowing, and second to demonstrate to the same layman how the scientific method has reached those conclusions. Such aims would seem to be mutually incompatible Elementary courses even in a single science are expected to convey so much information and so much basic doctrine that almost without exception they are authoritarian in presentation, and the student who continues in science learns that science is principally inquiry and doubt only at a later stage, commonly in graduate school if at all. On the other hand, courses of the "cultural" type, intended to introduce the layman to the scientific method, ordinarily select small segments of science which are used to exemplify the method, and no attempt is made to present the full sweep of scientific discovery even for one major field. But Krauskopf has not been content to discuss only simple and limited portions of the physical sciences, but has tried to cover all their major aspects, including such diverse topics as colloid chemistry, air-mass analysis, and the general theory of relativity, manifestly expecting the student to understand each of them, and he has at the same time restricted himself in mathematics to the simplest arithmetic and algebra and an occasional graph. How then can he hope to give the student even an inkling of the real meaning of the scientific method?

The extraordinary thing is that he has succeeded. There is hardly a major conclusion of physical science today that is not at least touched on in this book, yet the student is made constantly aware of the tentative groping process that led to those conclusions, he is introduced not only to the fields where the scientific method has made its great successes but also to those where it is still cautiously feeling its way. In the reviewer's opinion, two characteristics of the book have made this possible. The first is the historical approach, which is used in subject after subject, instead of first presenting the magnificent simplicity of the generalizations and then showing how well they explain the facts, Krauskopf has tried first to present the bewildering complexity of the facts and then to show how certain gifted individuals, many of them mentioned by name, piled inference upon inference until they arrived at general "law." In so doing he has cut across the conventional divisions between the sciences whenever the argument required, and has grouped the subjects treated into six sections, centering around the concepts of gravitation, the atom, the electron and the nucleus, chemical reaction, earth history, and the universe beyond the earth.

The second characteristic of the book is scientific humility, the constant admission that science and its method are not all powerful but proceed by trial and error, succeeding in one field but failing to advance in another, and the constant emphasis on the concept that even science's most cherished doctrines are either postulates or inferences which may require modification in the future as such postulates as "no action at a distance" and such inferences as the law of the conservation of matter have required modification in the past. As a result, the serious reader sees rising before him the whole structure of scientific hypothesis, not springing full-grown like the goddess Athena, but being put together piece by piece through the patient efforts of many men.

If the reviewer has any qualms about the book, it is precisely because of certain deviations from the two characteristics mentioned. Thus the historical approach is so successfully used for the theory of gravitation that one wonders why the electron theory of electricity was introduced ahead of the discussion of electric currents. Again, the inductive approach is occasionally abandoned for the authoritarian, and oversimplifications such as the neat separation of the geologic eras by catastrophic revolutions creep in. Probably however such deviations were enforced by the necessity of crowding the immense variety of information treated into a book of reasonable size and cost. One is tempted also to suggest alternative arrangements of the subject matter, but as the present arrangement was worked out on the basis of practical teaching experience, it is not to be tampered with lightly. As it stands, the book is an extraordinarily comprehensive review of physical science from which, however breathless he may become as he jumps from subject to subject, the student of reasonable intellect and interest cannot fail to attain an appreciation of both the results of science and its method.

JOHN RODGERS

Crystal Structures; by RALPH W. G. WYCKOFF. Pp. 491. New York and London, 1948 (Interscience Publishers, Inc, $8 00; looseleaf binding, 8 5 x 10.3 inches.) There can be no question of the need for, or value and usefulness of, a new edition of Dr. Wyckoff's well-known compilation of the structures of crystals. The pagination in no less than twenty-seven groups will at first seem a considerable objection to the new edition, but this feature is a small price indeed to pay for the page replacements containing additions or corrections, promised in the Publisher's Note at the beginning of the book. By placing the chapter number and title at the top of every page and providing an extensive table of contents, adequate means is provided for finding the descriptions of crystal structures.

The plan of the book is similar to that of the previous editions: Chapter I, Preface and introduction; Chapter II, Structures of the elements, Chapter III, Structures of the compounds RX; IV, RX_2; V, R_mX_n; VI, $R(MX_2)_n$; VII, $R_n(MX_3)_p$, constitute the present section I of the book. Section II, Chapters VIII through XII, will cover the remaining inorganic crystal structures including silicates; and organic structures will be described in Section III.

Sections II and III together will have to be somewhat smaller than Section I, to be accommodated within the looseleaf binder furnished with Section I. Dr. Wyckoff may find it hard to confine his expanding book to that space, but few will object to an additional looseleaf binder if and when it becomes necessary

This monumental work is a necessary reference for all crystallographic workers. Crystallographers owe a great debt of gratitude for Dr. Wyckoff's efforts in assembling it.

HORACE WINCHELL

A Source Book in Greek Science; by MORRIS R. COHEN and I. E. DRABKIN. New York, 1948. Pp. xxi, 579 (McGraw-Hill Co., $9.00). The need for our enlivening the traditional teaching of science courses by historical perspective has often been emphasized. Unfortunately, the task of organizing the material required for that task had proved too large and time-consuming for working scientists. The publication of "A Source Book in Greek Science" has eliminated this obstacle.

For it presents between the covers of a single volume most of what is likely to be regarded as interesting and useful in Greek literature by the student of the natural sciences. The selections are well made. Some of the passages appear in English translation for the first time. The book is not a collection of fragments; it presents the story of each science in fairly coherent form, the connecting material being skillfully provided by the authors where-ever the original requires an introduction or an explanation to fill a gap. Particularly noteworthy is the modern outlook which asserts itself not only in the arrangement of the historical material but in the footnotes and in the connecting passages as well.

Physics gets the lion's share of attention, and it needs it perhaps more than the other sciences because of the indifference and, to a large extent, misrepresentation which Greek physics has suffered in its treatment by physics teachers. A good deal of Aristotle is included, and it is no longer defensible in view of the publication of this volume to reiterate the textbook judgment which condemns Aristotle as having a non-scientific approach. Next in extent of treatment are medicine and biology, mathematics, astronomy and chemistry fill the remainder.

This book will find its way into all scientific libraries and, we hope, into the libraries of many science teachers.

HENRY MARGENAU

PUBLICATIONS RECENTLY RECEIVED

Organic Syntheses. Vol. 28; edited by R. S Snyder. New York, 1948 (John Wiley & Sons, $2 50)

Foundations of Modern Physics; by T B. Brown 2nd ed. New York, 1949 (John Wiley & Sons, $5 00).

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Handbook of Uranium Minerals. A Practical Guide for Uranium Prospecting; by J. De Ment, and U C. Dake. 2nd ed Portland, Oregon, 1948 (Mineralogist Pub. Co, $2 00)

Physical Geology and Man, by K. K Landes New York, 1948 (Prentice-Hall, Inc., $4 50).

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Procedure in Taxonomy; by A. Myra Keen and S W. Muller. Sanford, Calif., 1949 (Stanford University Press, $2 50)

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Illinois Geological Survey Bulletin No 72. Briquetting Illinois Coals without Binder; by R. J Piersol. Report of Investigations No 137. Viscosity Studies of System CaO-MgO-Al_2O_3-SiO_2 11, CaO-Al_2O_3-SiO_2; by J S Maehin and Tin Boo Yee. No 138 Colloid Properties of Layer Silicates, by W F Bradley and R E Grim Circular No. 148. Airplane Views of Illinois Oil Installations, by F. Squires Urbana, 1948.

A Brief Biology; by James Watt Mavor. New York, 1949 (The Macmillan Co, $4 00).

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American Journal of Science

JUNE 1949

THE LACCOLITHIC SERIES

O. T. JONES AND W. J. PUGH

ABSTRACT The paper refers to recent work on albite dolerites in an Ordovician area between Builth and Llandrindod, Central Wales. Near Llandrindod an intrusion formerly mapped as a single dolerite has been proved to consist of 315 separate intrusions each surrounded by Lower Ordovician shales. Most of these are small bodies of mushroom form, and exposures of two of them in quarries show that they were fed along narrow pipe-like channels or stalks at right angles to their bases. The form of the space within which these intrusions are contained is similar to an asymmetrical laccolith At Welfield near Builth dolerites are intruded as thin interrupted sheets at many different levels in shales; most are of flat mushroom form, but some are wing-like prolongations from short wide masses which are interpreted as feeders Some very small sub-circular intrusions may be feeders like those at Llandrindod. Here also the general form of the intrusive complex is that of a laccolith which shows certain resemblances to a cedar-tree laccolith. In the Western United States intrusions have been described which are transitional between the Welsh examples and Gilbert's ideal laccolith, and the authors regard all these as forms in a laccolithic series, individual examples being distinguished by their internal structure and relative amounts of igneous rock and sediments. Some remarks are made concerning Hunt's recent criticism of Gilbert's ideal laccolith.

INTRODUCTION

THE laccolith was first recognized as an important intrusive body about 70 years ago by Gilbert (1877). The typical laccolith of the Henry Mountains is a dome-shaped mass of igneous rock intruded into horizontally disposed sedimentary rocks, which are arched up by the intrusion, but the base of the intrusion is believed to be flat. An essential feature of this type of intrusion is that the sedimentary cover of the laccolith at any given point is arched by an amount which is equal to the thickness of the igneous rock below that point. Gilbert estimated that the sedimentary rocks which formed the arched cover of the laccoliths may have been from 7,000 to 11,000 feet in thickness and he believed that a considerable thickness of cover was necessary for intrusions to assume a laccolithic form.

A thin stalk at the base represents the igneous rock which congealed in the channel up which the magma travelled through the underlying strata. The rocks below the base of the intrusion retained their horizontal disposition and were unaffected except for the rupture through them of the channel which is now represented by the stalk. It has been said that the stalk was an invention of the artist-geologist, W. H. Holmes. Some years ago, one of us (O.T.J.) during a visit to Holmes at his Washington office took the opportunity of asking him if that was in fact the case. He said he thought something of the kind was required in order to provide a channel for the magma. No such structure has ever been seen in any of the Henry Mountains laccoliths, nor, so far as we know until recently, in any other comparable intrusive forms.

Gilbert recognized that there might be various kinds of laccoliths which departed from his ideal form. He interpreted some igneous masses as compound laccoliths, made of two or more intrusions, situated so close together that the arching of the sediments was due to the combined effect of the intrusions. Another modification commonly occurred where several discontinuous thin sheet intrusions were interleaved with sediments in the lower part of the cover. He also suggested that an intrusive mass, conforming more or less closely to the simple ideal form, might include within it a certain amount of sedimentary rock.

It was not long before other intrusive masses, which conformed more or less to the ideal form yet showing departures from it of various kinds, were recognized elsewhere in the United States of America and other parts of the world with the result that the concept of the laccolith gained worldwide acceptance. Recently there has been an attempt to cast doubt upon Gilbert's simple interpretation of the intrusions of the Henry Mountains and a new interpretation has been proposed by Mr. C. B. Hunt (1942) which differs in essential features from that of Gilbert; some comments are offered later upon this interpretation.

It may be recalled that the Henry Mountains are situated in a relatively stable region of the earth's crust and when the laccoliths were intruded the sediments were on the whole disposed in horizontal layers. Such a region is in strong contrast with the geosynclinal regions of the Cordilleras to the west or of the Appalachians to the east. Certain types of igneous activity appear to be closely associated with stable regions, other types are believed to be related to geosynclinal conditions and yet others to tectonically active regions. This aspect of the subject has been well discussed by Harker (1909) and others.

The work upon which we have been engaged during the past ten years in Central Wales appears to throw new light upon the structure of laccolithic bodies.

THE BUILTH-LLANDRINDOD ORDOVICIAN INLIER

The area which we have investigated is an inlier of Ordovician rocks occupying about 40 square miles and surrounded by unconformably overlying Silurian strata. We have mapped it on the scale of 1:2500 or approximately 25 inches to the mile.

The Ordovician rocks of the inlier and a nearby area consist in large part of dark shales, which can be sub-divided into five graptolite zones, namely, in descending order, *Dicranograptus*, *Nemagraptus gracilis*, *Glyptograptus teretiusculus*, *Didymograptus murchisoni* and *D. bifidus*. The first-named belongs to the Lower Bala, the next two to the Llandeilo and the last two to the Llanvirn, while in the area just outside the inlier Upper Bala grey mudstones also emerge from beneath the Silurian cover.

The base of the Ordovician is not seen in the inlier but by comparison with other areas it is inferred that the Llanvirn is underlain by the Arenig Series, mainly shales, and that series by the Cambrian System consisting probably of shales and sandstones or quartzites, which may be assumed to rest unconformably upon pre-Cambrian rocks. The thickness of rocks actually exposed up to the top of the Lower Bala varies from about 8,000 to 10,000 feet, it is estimated that the total thickness of sedimentary rocks from the base of the Cambrian to the top of the Lower Bala is about 16,000 feet and that they are mainly argillaceous sediments.

In one part of the inlier there is an important development of volcanic rocks, principally of spilitic and keratophyric types; these occur in the zone of *Didymorgraptus murchisoni*. In addition, several hundred intrusions of keratophyres and albite-dolerites occur, each of which is confined to certain restricted horizons in the shales.

The Ordovician rocks up to the end of the Lower Bala period were apparently little affected by earth movements, but after that they were subjected to powerful movements of various types. These latter movements were certainly completed before the Silurian (Upper Llandovery) rocks were laid down and there is reason to believe that they had at least begun and may have been largely completed before the deposition of the Upper Bala rocks. The intrusions, including even those which occur in the Lower Bala (*Dicranograptus* shales), had already been emplaced before these movements began and we believe that the intrusive activity was in some way related to the onset of these movements.

Throughout the early part of the Ordovician the inlier was a relatively stable region and had probably been so from Pre-Cambrian times onwards, but some miles to the west lay the Welsh geosyncline, the history of which has been described by one of us (O.T.J.). We believe that, prior to the intrusive episode which is described below, the stable region, probably extended for many miles to the east.

We have recently described two igneous complexes within the inlier which are characterized by numerous intrusions of albite-dolerite into Llandeilo shales. One of these is at the northern end of the inlier near Llandrindod Wells or Llandrindod as it is commonly called (1948,a); the other is at the southern end at Welfield about 7 miles south of Llandrindod and near Builth Wells, or more shortly, Builth (1946).

LLANDRINDOD COMPLEX

The old one-inch Geological Survey map which was published in 1850 shows a triangular mass of "greenstone" near Llandrindod. We have found that a large proportion of the area so mapped consists in fact of shales into which have been intruded numerous separate masses of dolerite. If an envelope line is drawn so as to enclose the masses which we have mapped, its course is closely similar to the boundary of the "greenstone" as shown on the Geological Survey map. The dolerites are intruded into shales of the *Nemagraptus gracilis* zone and as a result of subsequent earth-movements the rocks have been tilted in a general north-westerly direction at about 40°.

We found that the "greenstone," far from being one continuous mass, consists of no less than 315 separate masses, each separated from the other by shales (fig. 1). Approxi-

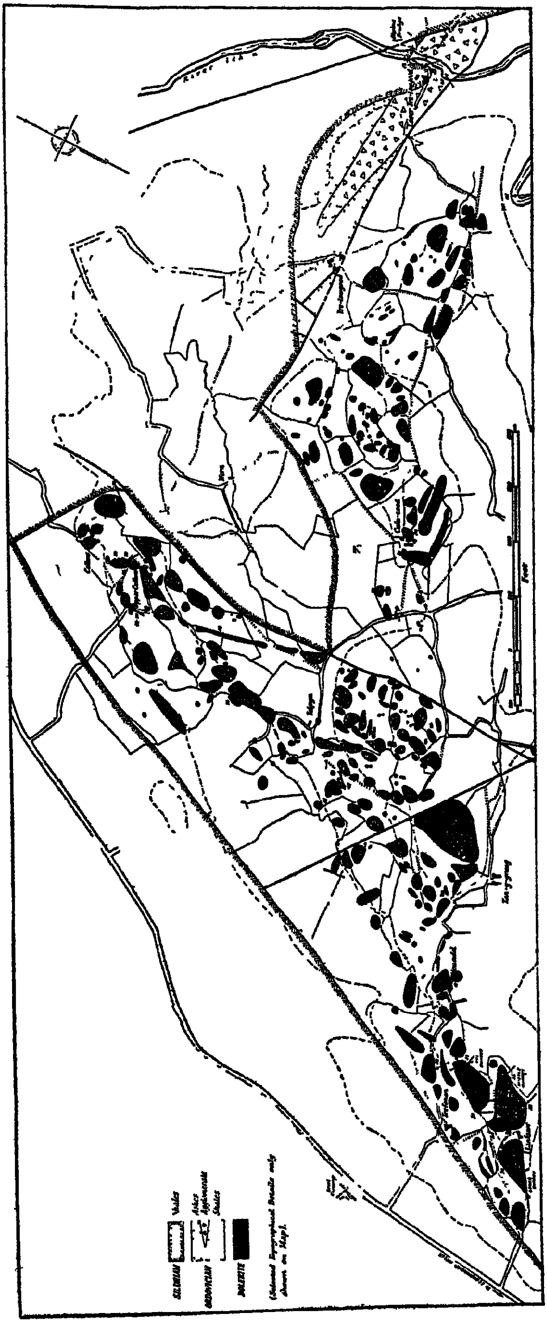

Fig 1. Map of dolerite complex near Llandrindod (dolerite masses black).

mately two-thirds of this number are oval or ovoid, but occasionally some are of more irregular form. Several of the individual dolerites have been quarried and their relation to the shales by which they are surrounded can be clearly seen. Quarry sections show that the bases of the ovoid masses are conformable to the shales below them and that each was emplaced in the shales in such a way that, while the base was relatively flat, the overlying shales were raised above the base by an amount which was proportional to the thickness of the intrusion. Each of these masses has, therefore, the form of an ideal laccolith but on a very small scale.

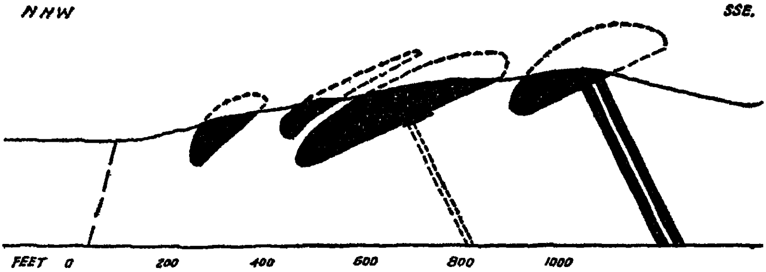

Fig 2 Section through dolerite masses at Llanfawr quarries, near Llandrindod.

Three large quarries have been worked at the southwestern end of the area near Llanfawr. The dolerite in the middle quarry is 600 feet long and has a maximum width of 300 feet. Quarrying operations have removed the igneous rock from above the shales for a distance of 120 yards and in doing so have uncovered two small masses of dolerite about 55 to 35 feet long by 18 to 16 feet wide, which project through at right angles to the dip of the shales, on which the quarried dolerite formerly rested Before the shales were tilted, these small dolerite masses were therefore vertical plug-like bodies; we were informed by the quarry foreman that they were joined to the base of the main dolerite and there seems to be no doubt that they were feeders to the masses above them (fig. 2).

When we first visited the upper quarry at Llanfawr there was visible at the northern end of the quarry a narrow bar of dolerite which projected downwards at right angles to the base of the main dolerite and reached the level of the quarry floor. When the lower end of this bar was exposed by excavation, it was found to end downwards in a sill about two feet thick. We were informed that it was first encountered about 30 feet back from the quarry face and its dimensions were therefore about 30 feet by 7 feet or rather more. Although its connection with the main mass was proved, one cannot say that it continued downwards below the thin sill, but it is not improbable that it did so (fig. 2). It is regrettable that this plug has now been largely quarried away but we were fortunately able to secure a photograph of it before it was removed, which is reproduced as a plate in our description of this area (1948, a, pl IV, fig. 2).

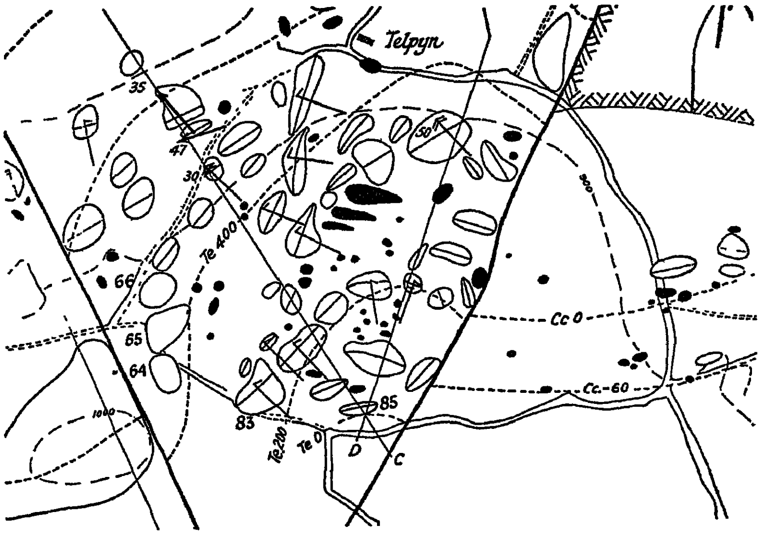

Fig. 3. Map of part of Llandrindod laccolith showing concordant dolerite masses (outline) and feeders (black) Straight lines within dolerite outcrop areas indicate direction of elongation of dolerite outcrops Single-headed arrow—dip of sedimentary rocks Double-headed arrow—dip of platy jointing in dolerite Single-barbed arrow—dip of dolerite mass as inferred from slope of ground Dashed line—approximate strike lines (in feet; Te—Telpyn block, Cc—Cefn-coed block)

Reproduced from part of Plate VI, Quart Jour. Geol Soc., 104, 1948, with permission.

We have thus found evidence in these two quarries at Llanfawr of three small oval plugs of dolerite which entered the main dolerite mass from below at right angles to its base. Their size in relation to that of the main mass is comparable with that of the stalk which is diagrammatically represented at the base of Gilbert's ideal laccolith. We have thus established that certain intrusions reproduce closely the form as well as some of the other characteristics of the ideal laccolith, though on a much smaller scale, and that these were fed from below through narrow channels or stalks.

Within the intrusive area at Llandrindod there are numerous small oval outcrops of dolerite which compare in size and shape with the plugs exposed in the Llanfawr quarries; others are almost circular (fig. 3). It is difficult to believe that such small masses could have been concordant intrusions and it is much more probable that most, if not all, of them were feeders to ovoid bodies at a higher level which have been removed by erosion. There are, in addition, some larger masses of oval or elongated form which are discordant in direction with other masses nearby and may also be feeders. They should probably be regarded as short dykes comparable with dolerite masses to be described later in connection with the Welfield complex. The total number of bodies which we infer to be feeders of one type or another is about one-third of the whole or over 100. It is thus probable that there formerly existed in the area at least 100 more concordant masses which have been removed by erosion, but by far the largest number of possible feeders are likely to be concealed beneath the surface.

The concordant ovoid bodies vary in length from about 600 feet to as little as 50 feet; their breadth is usually about two-thirds to three-quarters of their length and in a cross-section the height of the mass is therefore less than half the diameter. These masses tend to be arranged at successive stratigraphical levels through a vertical range of at least 800 feet of shales; this can be seen from the map although it is not immediately obvious. The complex has thus a distinctly layered structure; it is somewhat affected by small folds which cross it nearly at right angles to the general strike of the rocks in this part of the inlier and by two faults which break its continuity.

The former height of the structure is not known. If we consider the space or volume of ground within which the dolerites of the complex have been intruded, it is not unlike an asymmetrical laccolith in form. The base is not everywhere at the same horizon, but examples of laccoliths have also been described in which the base is transgressive.

It is because of the general resemblance between the space within which dolerites occur and the form of a laccolith that we assign this complex to the laccolith series. The essential feature which distinguishes it from the ideal form is that it is a layered structure composed of shales and discontinuous dolerite masses. We estimate that, as exposed in section at the present surface of the ground, the aggregate area of the outcrops of dolerite is only about 25 per cent of the total area defined by an envelope line drawn around the complex. It is, therefore, a very open structure. Moreover, not more than 25 per cent of the height of the domed form of the envelope can be attributed to the lifting of the strata by the intrusions; the other 75 per cent is to be accounted for by the fact that the layers of intrusions are emplaced at successively higher stratigraphical levels in the shales.

It is clear that the distribution of the ovoid masses, as well as the limits of the structure as a whole, were determined by the distribution of feeders. The height to which igneous material ascended in the shales depended possibly upon the size of the feeders and upon other factors, such as the viscosity of the magma and its temperature. At the time of the intrusive activity the shales probably contained a high percentage of water and we believe that most of the characteristic features of the Llandrindod laccolith are due to the fact that the magma was intruded into these wet shales probably beneath a relatively small thickness of cover (Jones and Pugh, 1948,b).

WELFIELD COMPLEX

We had mapped and described the Welfield complex (1946) before we examined the Llandrindod laccolith. The complex forms a prominent ridge north of the River Wye in which dolerites associated with highly baked shales are well exposed. On the west side, Silurian (Upper Llandovery) rocks rest unconformably on Ordovician dolerites and shales which belong to the *Nemagraptus gracilis* zone. These shales are estimated to be about 1,000 feet (or about 1,200 feet if the dolerites are included) higher in the stratigraphical succession than the rocks exposed in Welfield ridge, which are in the upper part of the *Glyptograptus teretiusculus* zone.

In Welfield ridge 44 separate masses of dolerite have been mapped but as the complex is truncated by a tear-fault on the north it seems probable that only about one-half of it is seen (fig. 4); the outcrops of most of the masses are oval. At least five of them, however, are different in form from any masses seen at Llandrindod; they are wing-like prolongations from two roughly oval dolerite outcrops which occur near the crest of the structure and are aligned approximately in the direction of its trend. The aggregate area of dolerite outcrops is roughly 50 per cent of the area included between an envelope drawn through the ends of the dolerites and the tear-fault which crosses the complex. From the limited information available the Welfield structure thus has a higher proportion of dolerites within it than the Llandrindod complex. This is in accordance with a rule generally applicable to the dolerites at different horizons within the inlier, which is, that those at lower levels in the stratigraphical succession are more massive and more continuous than those at higher levels.

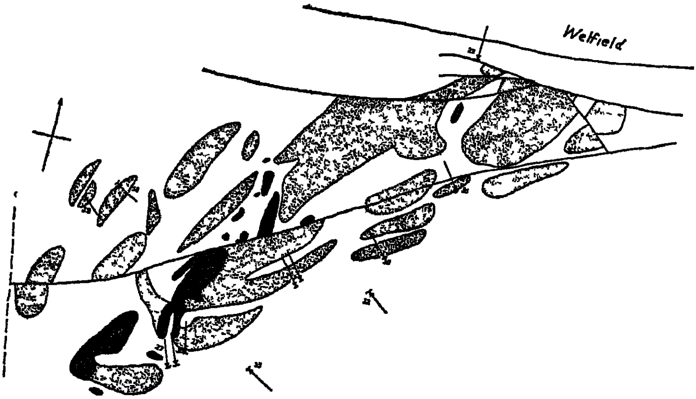

Fig 4. Map of Welfield laccolith showing feeders and possible feeders in black (other dolerites stippled).

The Welfield area has been affected by post-Silurian tilting and faulting. If allowance is made for post-Silurian tilting, the Welfield structure had the form of a laccolith with a dip in the western flank of 30° and in the eastern flank of 60°. At the present time the eastern flank dips at about 30° but, a short distance away from the strong feature which defines the eastern margin of the ridge, the Ordovician shales now dip in a westerly direction of 32° to 35°. These westerly dips when corrected for a post-Silurian westerly tilt of 30° become 2° to 5° and in combination with the dip of 60° in the eastern flank of the structure indicate a sharply defined flexure at the edge of the structure. They imply also that the Ordovician rocks at the time of the intrusive episode were almost horizontal. The corresponding differences in the dips on the western margin are probably only about 10° to 20° and the flexure is not so sharply defined. The existence of these flexures is regarded as an important confirmation of our view that the Welfield complex should be assigned to the laccolithic series.

We have no direct knowledge of the inner core of this laccolith. Since the present surface is a random section through the outer skin of the structure, it is probably fair to assume that the arrangement of dolerite and shale there seen is broadly similar to that which occurs below the surface. In other words, the shales at a deeper level probably include many ovoid intrusions with possibly some wing-like bodies given off from the downward prolongation of the dyke-feeders near the crest of the structure or from similar feeders which do not obviously play this role at the surface. In comparison with the Llandrindod complex it is not unlikely that several other masses which we mapped at Welfield may have been feeders also but at that time we had no reason to assume the existence of any type of feeder other than the two dyke-like bodies to which we called attention. These are shown in black in figure 4. We suggested that the unusual degree of baking of the shales may be due to larger bodies concealed beneath the surface.

COMPARISON OF LLANDRINDOD AND WELFIELD COMPLEXES AND OTHER INTRUSIVE FORMS

It is hardly necessary to recall the similarity between the Welfield complex and the Llandrindod complex. They are both layered structures and they differ only in two respects. The Llandrindod structure lacks the wing-like extensions found at Welfield and the latter contains more dolerites in proportion to associated shales.

In comparing the affinities of the Welfield complex with other intrusive forms we found that its nearest analogues were among the various types of laccoliths which have been described and figured from the Western Territories of the United States of America. We were impressed by the fact that

Welfield ridge or as much of it as can be observed revealed all the features that were considered to be characteristic of laccolithic intrusions. Moreover, the relation of the two short dolerite outcrops near the crest of the structure to the five wing-like masses which spring from them inevitably recalled certain parts of a cedar-tree laccolith.

The Welfield complex is a layered intrusion which stands in an intermediate position between the Llandrindod complex on the one hand and a cedar-tree laccolith on the other; and on this ground we assign both complexes to the laccolith series. They are both determined essentially by the incidence of feeders of one type or another.

It is hardly necessary to point out that the cedar-tree laccolith is one of the most striking examples of a layered intrusion that has yet been described. As represented by Holmes (1877), the individual layers all appear to be lateral or wing-like offshoots from a central mass which acts as their parent feeder; in comparison with the length of the wing-like extensions the central feeder is of considerable size.

Whitman Cross (1984, p. 209) has remarked that "as drawn, this intrusion implies a considerable absorption or assimilation of strata, for which no evidence is given and which is entirely improbable." It would be of great interest if this particular feature in the structure of the La Plata Mountains were re-examined and mapped in detail. In the light of our experience it may be found that the bodies which fed the leaves of the intrusion were less massive and more numerous than is indicated in the sketch by Holmes and that the appearance of incorporation of sediments suggested in that illustration may not be confirmed.

It is easy to envisage other intrusions of broadly laccolithic form in which the proportion of sedimentary rock to intrusive rock is considerably less than in any of the examples already mentioned. The el Late laccolith figured and described by Holmes (1877) and discussed by Whitman Cross (1894) appears to be a good instance. It is worth noting that Holmes was examining this and neighbouring areas at the same time as Gilbert was exploring the Henry Mountains. It is not improbable that the broad concept of the laccolith was already in the air at that time although it did not receive its name until 1877. It may be mentioned also that Holmes' sketches and in particular the figures of the el Late structure show that he had correctly envisaged the relation of the arching of the strata to the thickness of the underlying intrusion. This is clearly seen in fig. 5, which reproduces Holmes' sections of el Late; the first of these is well-known but the other two may be less familiar.

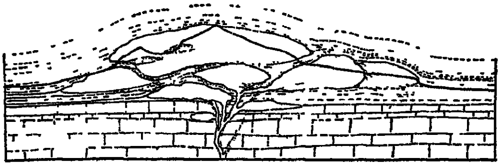

Fig 1
Section showing probable method of intrusion of masses of trachyte.

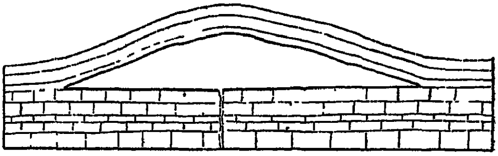

Fig 2.
Arching of strata produced by intrusion of single mass uniformly distributed.

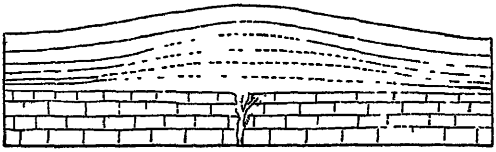

Fig 3.
Am. Photo Litho Co. N.Y. (Osborne's process)
Degree of arching really produced by the irregular intrusions

Intrusion of masses of Trachyte.
Sierra el Late.

Fig. 5. Sections of Sierra el Late by W. H. Holmes (1877).

In the el Late structure the intrusive bodies are separated by relatively narrow partitions of stratified rocks which, so far as can be judged from illustrations, occupy only about 25 per cent as compared with 75 per cent in the Llandrindod complex. No one has any difficulty in accepting the el Late structure as a variant of the ideal form and in regarding it as a compound laccolith. In the el Late laccolith about 75 per cent of the total amount of arching was directly due to the intrusive rock and only 25 per cent to its layered structure as is illustrated in Holmes's figures but it is clear that the masses in the upper part of the complex were intruded into a higher level of the stratified series than those near the base of the laccolith.

It is unnecessary to discuss the many other types of laccoliths which have been described and are transitional between the el Late laccolith and Gilbert's ideal form in which the whole of the arching of the covering strata is due to the intrusion. If, however, the examples mentioned above are considered together it is possible to recognize every transition between, on the one hand, a much divided structure such as the Llandrindod complex and, on the other hand, Gilbert's ideal form. This is the reason why we assign the rather remarkable intrusive forms displayed at Welfield and Llandrindod to a laccolithic series. Their underlying unity is the form of the space in which the intrusive rock is included and this as in the ideal laccolith is determined broadly by the incidence of feeders. The detailed arrangement of the individual masses which make up the whole structure depends upon many factors. But as Harker (1909, p. 70) says, "In the case of the ideal laccolite intruded in connection with crustal stresses purely of the plateau-building kind, the precise location of the intrusions will depend upon intratelluric conditions beyond our cognisance." So far as the Welsh examples are concerned, we have discussed some of these questions in the publications already mentioned.

COMMENTS ON A NEW INTERPRETATION OF THE HENRY MOUNTAINS LACCOLITHS

Charles B. Hunt (1942) recently claimed that the large symmetrical domes of the Henry Mountains are not due to mushroom laccoliths but have resulted from the injection of stocks. This claim rests not upon actual proof of the internal structure of these mountains but upon certain theoretical considerations. According to Gilbert, these mountain domes resulted from the intrusion of a magma at a certain horizon which lifted up the strata above that level into a dome the height of which depended upon the thickness of the intrusive mass. It was to the igneous rock of this general form that the name laccolith was originally applied; with it were associated sheets of sill-like form, and dykes.

Hunt disguises the issue by re-classifying the igneous rocks into "stocks, laccoliths, bysmaliths, dikes and sills" (p. 197) but the bodies termed laccoliths are defined as "elongate, tongue-shaped masses that were injected radially into the strata adjoining the stocks" (p. 197). These would appear to correspond in general to Gilbert's sheets and are unessential accompaniments of the main body of the laccolith; their presence or absence has no bearing upon the form of that body. By this somewhat arbitrary re-definition of the term laccolith which seems to rest upon no solid basis, Hunt is able to call by the name of stock what everyone has hitherto accepted as *the* laccolith.

In the very small-scale cross-sections (fig. 1, p. 200) which illustrate the paper there is no indication of the horizontal layers which occur at a lower level than the base of Gilbert's laccolith. These horizontally disposed layers are prominently displayed in several of Holmes' sketches but they have no place in Hunt's theory of these domes. He states that "the stocks of each of the five mountain masses must have been emplaced by physical injection because the space occupied by the stocks would be closed if the formations turned up around the stocks were restored to their original horizontal position" (p. 198). This appears to state no more than that the intrusion made room for itself by pushing away the surrounding strata and applies just as well to the ideal laccolith as to the supposed stocks.

The main argument seems to rest, however, upon the statement that the width of the stock, that is, the exposed part of the igneous mass near the centre of the dome in each of the four mountains, Mt. Holmes, Mt. Ellsworth, Mt. Pennell and Mt. Hillers, "is almost a direct function of the amount of uplift" (p. 198). These are compared with the Navajo Mountains on the Colorado Plateau about 50 miles to the south, which is of about the same diameter at the base (6 miles) as each of the above domes. The Navajo dome uplift is estimated to have been about 2,600 feet; it is calculated that the rocks within the basal perimeter which had an area before uplift of 9π or 28.27 square miles must have occupied after uplift an area on the surface of the dome of 28.75 square miles, indicating an areal increase of 0.48 square mile or 1.7 per cent. It is not clear how this figure was obtained. If the uplift was approximately in the form of a spherical cap, the increase of area where the height is one-sixth of the radius of the base, as in this case, should be 2.78 per cent or nearly 0.8 square mile. Hunt's diagram (fig. 1) represents the uplift as conical in form, but if this form is assumed the increase in area with the above ratio is about 1.4 per cent or about 0.4 square mile. Perhaps he has adopted some compromise between these extreme forms between which would lie most dome-like uplifts. It is not, however, with the precise amount of the areal increase that we find difficulty but in the argument on which it is based and on the use which is made of it.

In the Navajo dome, according to the author, the stretching by uplift has not been so great that the strata at the summit of the dome have parted so as to allow of igneous intrusion. The increase in area of half a square mile is thus assumed to give a limit to the amount of stretching that can occur in the cover-rocks without rupture, but the validity of this argument is very doubtful. If, for example, the dome were composed of a rock with a modulus of elasticity of about 5×10^6 lbs. per square inch, such a rock would rupture under a linear strain of about 1/2500 inch per inch. This is not very different from what might be caused by simple thermal expansion. The areal stretching just prior to rupture would be 1/1250, which is only about one twentieth of that attributed to the formation of the Navajo dome. It is obvious, therefore, that the increase in area of the rocks of Navajo dome could not be due to simple stretching but must have been accomplished by rupture, probably by the opening of joints. This process once started can go on to an unlimited extent all over the domed surface but possibly more near the crest than down the flanks and, therefore, can give no basis for the assumption that the rocks can stand so much stretching but no more without opening up.

It is then argued that "greater uplift in areas of the same size in the Henry Mountains has resulted in the parting of the strata and the creation of space to accommodate the stocks" (p. 199). In each diagram of fig. 1 (p. 200), the stippled hypothetical area available for the intrusion of each stock has been reduced by an area equivalent to that represented by the stretching of the strata as determined at Navajo dome. Once the basis of the stretching hypothesis has been proved to be wrong it does not follow that the intrusion of a cylindrical plug-like form is necessary in order that the above relation shall hold. They can be accounted for equally well by assuming Gilbert's ideal laccolithic form, for it is the mere shape and size of the dome that determines these relations and not what is inside it.

There are three features of the Henry Mountains laccolith that tell strongly against this author's interpretation, namely, (1) the existence of horizontal strata below the general level of the inferred base of the laccolith, (2) the distance away from the igneous core to which alteration of the strata has been observed in some cases and (3) the relation of the cover to the igneous core, for example, in Mount Hillers and in Ragged Mountain described by Whitman Cross (1894, p. 181, fig. 29).

We may also add that in our view the injection of a cylindrical plug-like stock must take place by a process of shearing analogous to the driving of a punch into metal; this is accomplished by pushing away and compressing the adjacent material and the effects diminish rapidly as the distance from the shear zone increases. It is therefore most unlikely to give rise to a broad, uniformly curved doming of the strata such as occurs in the Henry Mountains.

According to Hunt, the strata around the Mount Holmes igneous core are metamorphosed and shattered for as much as a mile from the walls of the intrusion. In our experience, this indicates a high probability that the igneous rock is at no great distance below the surface for at least that distance from the exposed margin of the intrusion. In Mount Ellsworth, a narrow tongue-like belt of altered rocks about a mile and a quarter long is represented which likewise suggests the prolongation of the upper surface of the intrusion to at least that distance.

Finally, Hunt appears to agree with some of the arguments presented above, for he remarks that "although the evidence shows that the stocks in the Henry Mountains were forcibly injected, the large domes around the stocks are not necessarily the result of that injection" (pp. 199-200). After considering four hypotheses to account for them and rejecting two which obviously do not apply, he concludes that they may be due either to arching by one huge, deeply buried laccolith of mushroom form or to a vertical push as if by the stocks. He objects to the first of these hypotheses on the ground that it would necessitate the assumption of an almost perfect symmetrical mushroom laccolith (*the* laccolith of Gilbert), whereas the known laccoliths (of Hunt or sheets of Gilbert) are tongue-shaped, elongate bulges. He remarks, moreover, that the assumed mushroom laccolith would have to be 20 times larger than the largest known laccolith (of Hunt, not Gilbert) in the Henry Mountains and about 50 times larger than the average volume of all known laccoliths (again of Hunt) in these mountains. Here we have examples of the confusing effects of the author's attempt at restricting the name laccolith to certain forms arbitrarily selected by himself.

He considers also that the growth of a laccolith to such huge size would probably result in imperfect form. We would comment that in the absence of directional stress a symmetrical form is the most likely to be produced.

A paragraph near the end seems to explain why the above suggestions are made by the author, "Although it is perfectly true that the domes could be the result of arching over laccoliths (? of Gilbert), they could as readily be the result of doming by stocks like those (assumed to occur) in the Henry Mountains. The object of this paper is to stress the need for reconsidering the explanations hitherto applied to large symmetrical domes associated with igneous intrusives." (p. 203.) The words in parentheses in this quotation have been inserted by us.

Our experience in Wales leads us to believe the truth of Gilbert's magnificent conception and we have also been fortunate enough to discover the stalks through which masses of laccolithic form, though on a very small scale, were fed.

Finally, Hunt refers to explorations by boring in the laccolith of the Marysville Buttes described by Howell Williams (1929). Although the original intrusion of the Marysville Buttes is regarded by Howel Williams as a steep-domed laccolith it has been considerably affected by subsequent volcanic and intrusive activity so that its original form has been almost destroyed. It is perhaps significant that a volcanic vent was later opened up just where one would assume the stalk of the original laccolith to have been. Is this accidental or may it give a hint of the existence of a stalk situated centrally below the original laccolith?

Hunt mentions several borings which were carried down for several thousand feet on the flanks of the structure; one of these was within a few hundred feet of the exposed outcrop of the igneous core and was carried to a depth of 7,014 feet. This did not reveal a continuous mass of igneous rock but a number of sills intruded into the domed strata. If a similar boring were put down, say, into the core of the Welfield laccolith, we should expect, in view of our present knowledge of these structures, to find numerous thin intrusive sheets interleaved with sediments and it may be that a similar state of affairs may occur in connection with laccolithic bodies more frequently than has been supposed.

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METASOMATIC GRANITIZATION OF BATHOLITHIC DIMENSIONS

PART II. STATIC GRANITIZATION IN SHEKU AREA, NORTHWEST YUNNAN (CHINA)[1]

PETER MISCH

ABSTRACT. Near Sheku in Northwest Yunnan, red beds deposited after late Triassic orogeny and barely affected by later Mesozoic orogeny which is important in neighboring parts, have been statically granitized within an area measuring 13 by 20 miles as mapped. The rocks can be described as "granodiorite-porphyries." They occur in irregular bodies formed entirely by metasomatic transformation of sediments. The virtually flat-lying red beds enclosing the granodioritic bodies, as well as relict inclusions of red beds in these bodies, have not been displaced. Some of the stratification of the red beds seems to have been preserved in the granitized rocks. The groundmass of the "granodiorite-porphyries" has the clastic texture and composition of the fine-grained quartzose-argillaceous red beds bordering the igneous-appearing rocks. Apparent phenocrysts of calcic oligoclase, microperthite, quartz, and subordinate hornblende and biotite are porphyroblasts showing replacement features. There are all stages from early crystalloblastic growth to "euhedral" porphyroblasts imitating phenocrysts Apparent resorption embayments in quartz have formed by differential crystalloblastic growth

The essential material added was alkali, with sodium predominating. The amount was small. Introduction was by rising solutions which also conveyed heat. Metasomatism occurred at an unusually shallow depth, and at a moderately high temperature Recrystallization of the red beds near the granitized bodies was incomplete, restricted, and predominantly at low temperature. In contrast to most areas of static granitization, there was no regional heating and recrystallization of the sediments outside the granitized bodies. This feature, the general preservation of clastic relict texture in the transformed rocks, and their porphyritic-appearing texture are attributed chiefly to the shallow depth The Sheku type of static granitization is compared with the "normal" type which in large transformed bodies produces "eugranitic"-appearing textures Amount of introduced material, and removal of some of the original material versus increase of volume, are discussed.

INTRODUCTION

THE general problem treated in the present paper has been stated in the introduction to the first part published in the April issue of this JOURNAL. In the first part, it has been shown that synorogenic granitization is capable of producing vast areas of gneiss-granitic rocks. The intimate relations between orogenic regional metamorphism and synkinematic granitization have been discussed and certain general conclusions suggested. After orogeny in geosynclines batholithic masses of granitic rocks are widely formed under static conditions. Do they owe their origin to magmatic intrusion or to metasomatic transformation of pre-existing rocks or to both? Among examples which I have had an opportunity to study in the field, an area is here described in which unusually favorable conditions permit of establishing a metasomatic origin of entire granitic masses, and not only their marginal portions. Then, this case is compared with other types of static granitization. For a recent review of the history of the ideas involved in granitization, I again refer to the 1948 Granite Memoir of the Geological Society of America, especially to H. H. Read's contribution.

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[1] Read during Geological Society of America, Cordilleran Section, 1948 meeting at Pasadena

I. OUTLINE OF GEOLOGY OF SHEKU AREA, NORTHWEST YUNNAN[2]

The area described is in the geosyncline which I found to occupy Western Yunnan, near the southwestern border of China (Misch, 1945 b). The area is west and south of Sheku, a town located on the bank of the upper Yangtsekiang where this river abruptly changes its southern course and turns back in a tremendous sweep. A geologic reconnaissance map of the area is given in plate 9, and a general section along the main traverse studied, in figure 1.

In this area there are three main groups of rocks: (1) *intensely folded and in part dynamically metamorphosed sediments and volcanic rocks ranging in age from pre-Cambrian(?) to Triassic*, (2) *a thick series of unconformable Mesozoic red beds*, and (3) *granitic rocks later than the red beds.* The sequence of formations composing the first group is outlined in the legend of the map (pl. 9). The granitic rocks treated in the present paper come on the east partly in contact with the oldest formation in this area, the presumably later pre-Cambrian *Sheku schist* (cf. Misch, 1945b, p. 51-52). It consists of thick micaschists grading into phyllites and is "upper"-mesozonal and "low"-epizonal. This formation, and *Paleozoic and Triassic rocks* which in most parts are epizonally metamorphosed, are involved in strong folding and complex thrusting which includes movement of older over younger as well as of younger over older rocks (cf. Misch, 1945a, p. 541-543). The latest fossiliferous rocks involved in this orogeny are of *early Norian* age (*Paratibetites adolphi*, etc., cf. Misch, 1947, p. 109-110). The western part of the area (cf. pl. 9) is composed of Triassic red beds and marine sediments associated with thick porphyries and rhyolites. These rocks are steeply folded. *Cardita crenata* indicates that Carnian is represented.

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[2] My field work in northwest Yunnan extended from 1939 to 1946, under the auspices of the National University of Peking which during that period was incorporated in the National Southwest Associated University at Kunming, Yunnan, and of Chinese Government institutions The field work in the Sheku area was carried out in 1940 and 1945. I wish to thank Prof Y C. Sun, Director of the Geology Department at the National University of Peking, for his permission to use for this paper the material collected in the field Most of the petrographic work was done in the Geology Department at the University of Washington

All these rocks are with strong angular unconformity overlaid by a thick succession of continental red beds, including shales, fine-grained sandstones, grits and conglomerates. This formation belongs to the writer's "upper red beds" of Yunnan (Misch, 1945b, pp. 98-102 and 145-148, and Misch, 1947, pp. 106-109).[3] In several areas the "upper red beds" are conformable with partly fossiliferous (lower) Norian black shales and gray sandstones. There the "upper red beds" include Rhetian, if not upper Norian, though they presumably extend far into the Jurassic. In the region here described, *intense Late Triassic orogeny* has taken place before the deposition of the "upper red beds"(cf. Misch, 1945b, p. 148), and it is therefore uncertain whether they also here began to form during the latest Triassic, or whether deposition did not begin till Jurassic time. A Jurassic age of the main part of the "upper red beds" seems to be confirmed by the discovery, in the southern part of the area shown in plate 9, of an intercalated limestone member in which a few supposedly Jurassic marine fossils were found. The thickness of the "upper red beds" west of Sheku is over 10,000 feet, to which has to be added an unknown thickness removed by erosion.

Late Triassic orogeny was intense in the area here described, but was found to be lacking in wide parts of the West Yunnan geosyncline. In most parts of the geosyncline, it was succeeded by equally intense *later Mesozoic orogeny* (cf. Misch, 1945a, and 1945b, pp. 119 and 122-123). "Upper red beds" were also involved in epizonal dynamometamorphism and overthrusting. However, in the area here described, this second orogeny is almost lacking. Most of the "upper red beds" are practically

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[3] I have classified the Mesozoic red beds of Yunnan into "lower red beds" = Lower Triassic, "middle red beds" = Middle and lower Upper Triassic; and "upper red beds" = highest Triassic and, in large part, presumably Jurassic. The first two divisions I have shown to interfinger laterally with marine sediments.

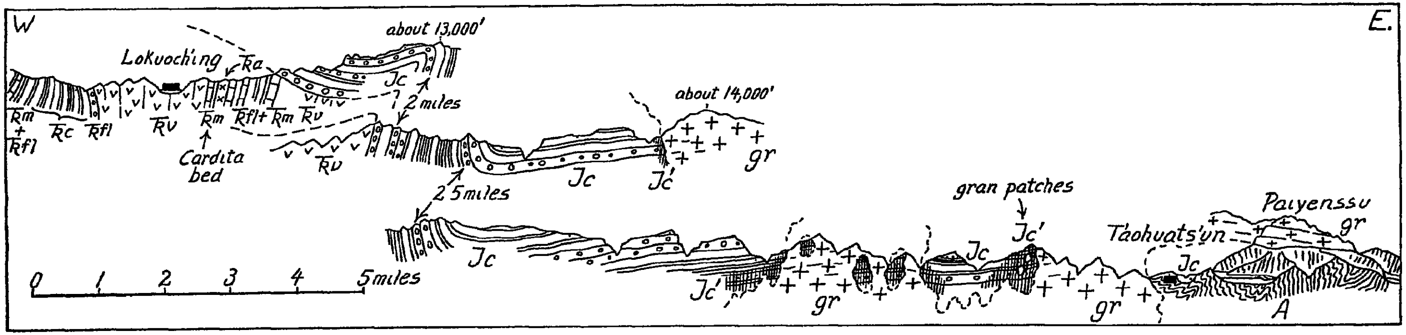

Figure 1. Section west and southwest of Sheku, Northwest Yunnan. Location of sections is marked on the map (pl. 9)

gr : "granodiorite-porphyry"
Jc : "upper red beds," Jurassic (and highest Triassic?)
Jc' : the same, more or less recrystallized
Tr v : rhyolites and porphyries, presumably Triassic
Tr a : Triassic andesite
Tr m : marine Triassic limestones and shales, including Carnian with *Cardita crenata*
Tr fl : Triassic flysch, gray sandstones with shales and conglomerates, partly continental
Tr c : "middle red beds," Middle and lower Upper Triassic
A · Algonkian micaschists

PLATE 9. *Geological Reconnaissance Map of Part of Northwest Yunnan (China)*

Q : Quaternary (smaller areas omitted)
Pc : Pliocene, lake beds and some gravels
Ta : Tertiary biotite-andesite
KTv : dacitic and andesitic flows, Cretaceous or Tertiary
gr : metasomatic "granodiorite-porphyry," Jurassic or Cretaceous; with local diorite
grp : pegmatite connected with gr
Jfe : felsite and related rocks, (?)Jurassic; probably transformed Jc
Jc : "upper red beds," Jurassic (and Rhetian ?)
Jc'' : the same, more or less incompletely recrystallized
JTr c : "middle" and/or "upper red beds"
JTr' c : the same, epizonally metamorphosed (phyllites, etc)
Tr cp : calc-phyllite, Upper Triassic; with quartz-dioritic dikes made into mylonite-schists
Tr$_3$: Norian black shales and gray sandstones (marine with continental intercalations)
Tr'$_3$: the same, epizonally metamorphosed (black phyllites, etc)
Tr$_2$: Ladinian and Carnian massive limestone (continuous in eastern facies; forming lenses and bands in western facies)
Tr fl : flysch; black shales and gray sandstones, continental and marine; probably including some Tr$_3$ . . . . } lateral equivalents of Ladinian-Carnian limestone in western facies
Tr c : "middle red beds" . . . . . }
(Tr ?)ls' : foliated limestone with some calc-phyllite; equivalent to Tr$_2$?
Tr a : Triassic andesite
qd : quartz-diorite-porphyry, Triassic or later
Tr v : rhyolite and porphyry, presumably Triassic
Tr v' : the same, epizonally metamorphosed (schistose porphyroid)
Tr$_1$: { Anisian shaly limestone . . . . } lower Middle and
{ Scythian red beds ("lower red beds") } Lower Triassic
uPm : Upper Permian marine shale
Pm v : Permian basalt (with intermediate and ultrabasic flows)
Pm v' : the same, epizonally metamorphosed (greenschists, etc.)
Pm ls : Lower Permian massive limestone
Pals : Devonian-Carboniferous massive limestone (possibly including some Lower Permian)
Pam : the same, epizonally metamorphosed (banded marble, etc.)
O : Ordovician shale, phyllite, and quartzite
A : Algonkian (?) micaschist
Ar : Archaean gneiss
S—S' : sections in text fig. 1.

undisturbed. Only in the western portion of the area are they folded (cf. fig. 1), but their folding is much weaker than that of the older rocks. In the area described, the "upper red beds" do not show any trace of differential deformation or dynamic metamorphism.

Within an area measuring about 13 by 20 miles as mapped, the "upper red beds" contain *irregular bodies of massive granitic rocks.*[4] They show a definite tendency to keep to the red beds, and only in the east do they partially come in contact with the underlying Sheku schist. Since the granitic-appearing rocks are later than the "upper red beds," they are considerably younger than the Late Triassic main orogeny in this region. Whether they are also later than the second, e.g. post-"upper red beds," orogeny of West Yunnan is uncertain since in the region here described this orogeny has acted only weakly and locally, and has not affected the "upper red beds" where they enclose granitic masses (not, at least, at any localities I have visited). The granitic rocks were formed under static conditions and completely lack structures indicative of stress, either inherited, contemporaneous or superposed. The age of the granitic rocks is presumably Late Mesozoic. Exact stratigraphic dating is impossible since there are no definite Cretaceous rocks known in West Yunnan. Not far east of the region here described, thick conglomerates and breccias are widespread which are younger than all the rocks so far described. I have tentatively assigned an Early Tertiary age to these conglomerates (cf. Misch, 1945b, pp. 133 and 152). In the southeastern part of the area shown in plate 9, the "upper red beds" are unconformably overlaid by flat-lying andesitic and dacitic flows and tuffs which are possibly Cretaceous but more probably Tertiary.

II. FIELD EVIDENCE OF METASOMATIC GRANITIZATION[5]

How have the granitic masses acquired their position within the red beds? Traditionally, they would be termed intrusive, and the question considered answered, but *field relations do not show any indication of mechanical emplacement by magmatic intrusion.* As in many postorogenic granitic masses, the picture is one of the granitic rocks filling an irregularly shaped space formerly occupied by sediments, without these sediments being visible anywhere—as if someone had with a pair of scissors cut out certain parts of the sedimentary section in a haphazard manner and substituted granitic rocks. In the region here described, this situation is particularly striking, because *the red beds enclosing the granitic masses are undisturbed,* whereas in other areas the enclosing sediments have usually been folded prior to formation of the granitic masses.

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[4] The term "granitic" is here used purely descriptively and in its widest sense (including granitic *s. str.* and granodioritic, evengranular and porphyritic rocks), and chiefly based on appearance in the field.

[5] The term "granitization" is here used in a wide sense, as defined by F. F. Grout (1941, p 1540) according to whom it *"includes a group of processes by which a solid rock is made more like granite than it was before, in minerals, or in texture and structure, or in both."* The products of granitization are therefore not always granites, though granite is the final goal

The field relations in the Sheku region, as observed along the main traverse surveyed, are shown in figure 1. The almost horizontal position of the red beds excludes the possibility of mechanical emplacement of magma by doming. Locally an extremely gentle dip of the red beds away from the contacts was noticed. It cannot possibly be an indication of doming by intrusion, since it would take tilting many times stronger and much more widespread, to create the room needed to accommodate even the visible granitic masses. If those extremely slight and local dips are at all related to the granitic masses, they would indicate an upward movement infinitely small compared to the volume of the granitic masses. A very slight increase of volume during metasomatic transformation of sandy argillaceous sediments into granitic rocks could account for this (cf. chapter IV).

There are smaller and larger *bodies of red beds enclosed in the granitic masses.* Their bedding has remained practically horizontal, as in the red bed areas adjacent to the granitic outcrops. This complete lack of disturbance and disruption of the stratification of the sediments is remarkable. It is contrary to magmatic intrusion which would of necessity imply mechanical displacement of the sediments, irrespective of what mechanism of intrusion is suggested. The absence of doming has already been discussed. Nor can stoping have been the means of emplacement. For in this case, red bed bodies enclosed in the granitic masses would show evidence of sinking, which is not the case, and at least most of them would have lost their horizontal position. Generally speaking, it is difficult to conceive how fragments of wall rock could be enclosed in magma without being rotated and shifted. Nor is there any indication of an upward movement of the red bed bodies enclosed in the granitic rocks.

I conclude that no intrusive mechanism could account for the field relations observed. How then can the "mysterious disappearance" of the sediments from the space now occupied by granitic rocks be explained? The only possible answer appears to be that *the sediments have not disappeared at all.* They are still there, but *have by metasomatic processes been transformed into granitic-appearing rocks.* No great chemical change was needed to achieve this result. The sandy and silty shales and fine-grained argillaceous sandstones which predominate in the red beds laterally prolonging the exposed granitic masses, hardly needed more than an addition of some alkali to acquire a granitic composition (essentially sodium, as the igneous-appearing rock is granodioritic; cf. chapter IV).

Those granitic outcrops and specimens I have examined, *lack any trace of flow structure,* either megascopic or microscopic. As will be shown below, these rocks consist of a fine-grained groundmass and larger crystals among which are columnar hornblende and tabular biotite. Such rocks would be rather well suited to record magmatic flow—if they were of igneous origin. On the other hand, at some places there are *suggestions of sedimentary relict bedding* in the igneous-appearing rocks. The precipitous south face of Paiyenssu where the granitic rock is unusually well exposed (cf. fig. 2), shows a structure strongly suggestive of nearly horizontal bedding, and parallel to the stratification of the red beds which prolong the granitic rock on the west (cf. also fig. 1, lower section, east part). The lower slopes of Paiyenssu consist of steeply folded Sheku schists. They underlie the red beds in this vicinity. The south face of Paiyenssu seems to consist of flat-lying granitized red beds. They are topographically above the old schists and presumably overlie them normally.

Where the *contacts* between the granitic-appearing rocks and the red beds were observed, they are usually rather sharp. This might be taken as evidence against granitization, but it appears that replacement is apt to produce both gradational and sharp contacts, depending on whether the replacement "front" remained stationary after reaching its maximum extent, and also on local controlling discontinuities in the replaced rock, such as slickensided fractures and joints. That granitization is able to produce sharp contacts, was first pointed out by C. E. Wegmann (1935, pp. 326-327), and has since been confirmed by several geologists.

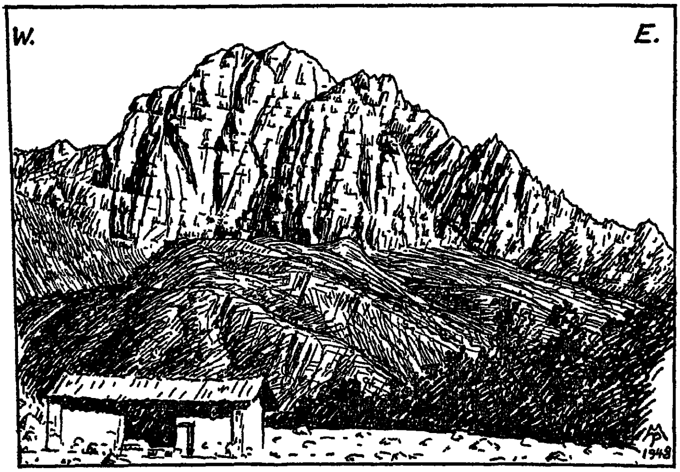


Figure 2 South face of Paiyenssu about three and a quarter miles northeast of T'aohuats'un, southwest of Sheku. Massive "granodiorite-porphyry" showing faint horizontal bedding. The soft lower slopes are composed of Sheku schist.

The interpretation of the granitic-appearing rock as a product of metasomatic transformation is further supported by *some small granitized patches* which were observed in altered red beds outside the larger granitic masses. A good example is exposed on the trail in the valley of T'aohuats'un (cf. fig. 1), and is shown in figure 3. The granitized patches have irregular outlines, and their contacts with the enclosing altered sediments are in part gradational and in part sharp, especially where control by joints is evident. These small granitized patches completely resemble the "small granodioritic blocks formed by additive metamorphism" described by G. E. Goodspeed (1937a) from Cornucopia in Oregon, and have formed in the same manner.

III. PETROGRAPHIC EVIDENCE OF GRANITIZATION

(a) *Description of altered red bed sediments*

Along their contacts with the granitic masses, the red beds have been altered to gray rocks of the appearance of fine-grained hornfels. The zone of alteration varies greatly in width (cf. fig. 1, and pl. 9). In most places it is narrow and locally it is not developed at all. Among localities showing a wide alteration zone, the high peak (about 17,000 feet) in the central part of the map (pl. 9) shows a peculiar section. Its lower south slopes are composed of unaltered red beds, which in its higher parts grade into gray altered rocks. The bedding is conformable from foot to summit, and probably the upper part entirely consists of altered red beds. Whether any granitized rocks are present, or whether they have been all eroded, is not known. More detailed work is needed than could be done during a rapid traverse, under the protection of a bodyguard of tribesmen armed with crossbows, of this bandit-ridden mountain range.

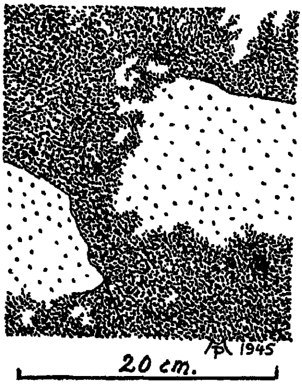


Figure 3. Replacement patches of "granodiorite-porphyry" in recrystallized argillaceous sandstone About two and a half miles west-southwest of T'aohuats'un, southwest of Sheku.

Under the microscope, the altered red beds reveal well-preserved *clastic textures*, and usually show a surprisingly *low degree of recrystallization*. Samples of dense-appearing to very fine-grained gray, brownish and greenish-gray rocks from the *western contact of the granitic mass in the valley of T'aohuats'un* and from *relict inclusions in the western portion of that mass* (traverse shown in fig. 1) show varying amounts of *small clastic quartz grains* set in a turbid matrix (pl. 2, fig. 1). The quartz grains are clear, and their shapes are angular to subrounded. Some of them have, due to beginning crystalloblastic growth, developed small projections and acquired amoebaelike outlines. The *matrix* is mostly turbid and stained by iron. Its constituents are variable amounts of *carbonate* and of presumably *kaolinitic material*, usually with *epidote*, some ill-defined *alkaline feldspar*, sometimes *sericite* and small amounts of *chlorite*. Epidote is rarely absent and its amount may be considerable. Most of it is present as irregular crystalloblastic patches lacking well-defined borders. Well-formed grains are occasionally found. Sometimes small projections of epidote are seen to penetrate clastic quartz grains. *Magnetite* forms tiny grains scattered in the matrix, or has concentrated in larger grains of irregular shapes. It is derived from the hematitic pigment of the original red beds. Apart from some ill-defined groundmass feldspar, there occur somewhat larger plagioclase grains, partly turbid and partly clear. Some of them may be clastic, and some newly formed. Occasional small flakes of muscovite and of pale brown biotite were observed. All these rocks are derived from fine sandy red shales and argillaceous fine-grained red sandstones, both calcareous, and laterally pass into these unaltered sediments. In a relict inclusion of altered red beds preserved in the granitic mass (fig. 1), a bed with megascopically visible small quartz pebbles occurs.

Samples of altered red beds from the *mountains south of the area just described*, include *similar rock types* (pl. 1, fig.1) and, in addition, *such showing a higher degree of recrystallization* which, however, still remains incomplete. Clastic grains of clear quartz are usually present. They are lacking in a few specimens derived from non-sandy argillites. *Epidote* and *sericite* are often present. Many samples contain amphibole. Pale green *actinolitic amphibole*, sometimes accompanied by a colorless amphibole, forms ill-defined crystalloblastic patches and feltlike aggregates, and rarely better developed larger grains. Dark green *hornblende* may occur in the same rocks, or independently. Its grains are usually better formed, and mostly larger. They occur singly, in patches, and compact aggregates. Sometimes, crystal form is approached. Often, green hornblende is segregated in microscopic veinlets, some of which clearly follow fractures. A few specimens are rich in brown *biotite*, which in one rock is accompanied by a large amount of idioblastic tourmaline. Most specimens contain *magnetite*. One specimen contains turbid sieve-textured *feldspar* *porphyroblasts* with crystal outlines. Another sample contains, in a poorly recrystallized extremely fine-grained groundmass, *composite patches of well crystallized hornblende and plagioclase.* Some of these patches have three- and four-sided outlines (pl. 7, fig. 2), and recall, on a microscopic scale, the "granodioritic blocks" described by G. E. Goodspeed (1937a) from Cornucopia, Oregon. Another specimen (pl. 6, fig. 2) has a finely brecciated sedimentary texture, containing angular quartz grains of various sizes some of which show strain shadows, and fragments of fine-grained quartzite or chert partly showing bedding. Some of the quartz has recrystallized as a mosaic free from strain shadows. Between the clastic particles, green hornblende has grown, and has partly concentrated in compact crystalloblastic aggregates. In this rock, all argillaceous matter and all carbonates originally present in the matrix have been used up to form hornblende. All other specimens examined still possess a *poorly recrystallized extremely fine-grained matrix,* locally with a "fluffy" appearance.

The generally rather low degree of crystallization and the mineral facies of the altered red beds described, show that the temperatures to which these rocks were subjected cannot have been very high. The contrast to true hornfels is certainly striking. An additional reason for the surprisingly low degree of crystallization and the small grain size of the newly formed minerals, is the complete absence of differential mechanical deformation in these rocks.

(b) *Description of granitized rocks* (*"granodiorite-porphyries"*)

The granitized rocks are massive and light gray to white. Their *appearance* in the field is *granitic.* They form high massive mountains and rugged faces, such as Paiyenssu ("White Stone Cliff") southwest of Sheku. *Megascopically,* the rock shows a medium-grained fabric of feldspar, quartz, and subordinate hornblende and biotite. Feldspar exceeds quartz. Large phenocryst-like quartz grains are common. They grade from rounded grains resembling small pebbles, to bipyramids. In the "interstitial" spaces, there is a megascopically very inconspicuous groundmass of light color. In a few places, varieties containing little or no quartz and relatively richer in mafics were observed. Although the field relations of these varieties have not been studied in detail, they are believed to belong together with the more acidic predominant rock type.

Microscopically, the feldspar grains are seen to be predominantly *calcic oligoclase*. Zoning is common, and is mostly recurrent. Frequently there are a dozen or more recurrences, and there may be as many as twenty. The composition of the plagioclase usually varies between An 25 and An 30. Occasionally there are cores of sodic andesine (An 31-32, An 33-34). Sometimes there are narrow rims of more sodic oligoclase (An 23, An 21, An 20). Otherwise, the variation of the anorthite content in individual zoned plagioclases rarely exceeds 4%, and commonly it is much less in adjacent recurrent zones. *Potash-feldspar* is associated with the plagioclase, forming grains of the same size. It commonly encloses plagioclase, is often intergrown with it, but also forms independent grains. Part of it is developed as microperthite. The amount of potash-feldspar is variable, but is usually considerably smaller than that of plagioclase. *Quartz* forms independent grains of various sizes. It frequently shows those embayments which are customarily attributed to "magmatic resorption"; it will be shown in chapter (c) that these are due to growth. The mafics are green *hornblende* and brown *biotite*. They do not attain the size of the feldspar and quartz grains. Hornblende is occasionally absent, but usually exceeds biotite. All these constituents vary from well formed crystals to irregular grains. Their textures will be described in chapter (c). Among accessories, *sphene* is usually present in comparatively large amounts, though occasionally it is lacking. *Apatite and magnetite* are generally present. Sometimes a small amount of *zircon* occurs. Occasionally some tourmaline is found.

All these minerals occur in a fine-grained *groundmass* consisting mostly of quartz and potash-feldspar, with some plagioclase. In most rocks the large crystals considerably exceed the groundmass, and therefore the megascopic appearance of the rocks is granitic or granite-porphyritic. In specimens from widely distant localities, and from marginal as well as interior portions of individual granitic-appearing outcrops, the groundmass consists of clastic-appearing angular and subrounded grains of clear quartz in a matrix of frequently turbid and sometimes iron-stained potash-feldspar (determined by its

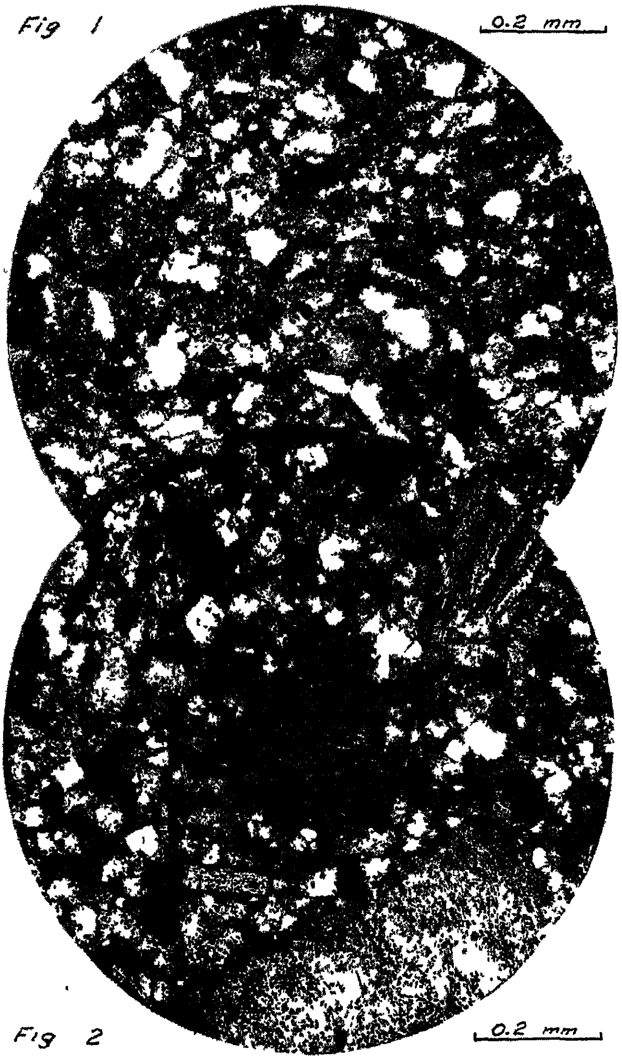
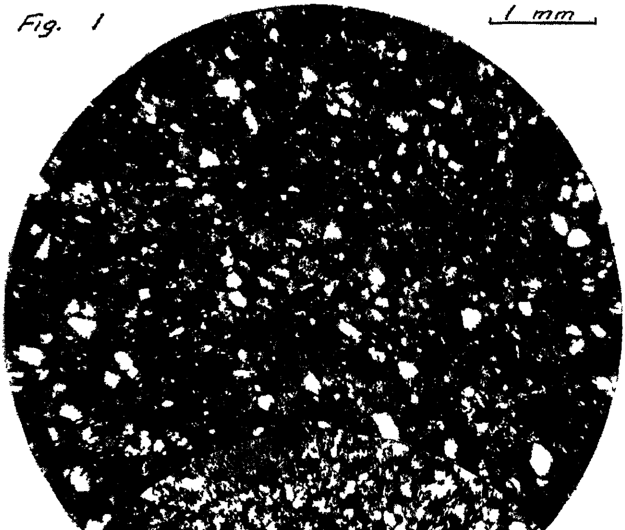


PLATE 1

Fig 1 Incompletely recrystallized fine-grained argillaceous red bed sandstone Clastic quartz grains, in part beginning to grow Groundmass green hornblende and quartz Crossed nicols South of upper part of T'aohuats'un valley, south of Hsiaoch'iaot'ou (specimen no 28 12 15 28)

Fig 2 Metasomatic "granodiorite-porphyry" Crossed nicols Groundmass of altered fine grained argillaceous red bed sandstone still shows clastic quartz grains Center incipient stage of crystalloblastic growth of replacement plagioclase Below part of a more advanced plagioclase porphyroblast Upper right hornblende porphyroblast Northwest of Chiench uan (no 22 1 16,1).

Fig. 1



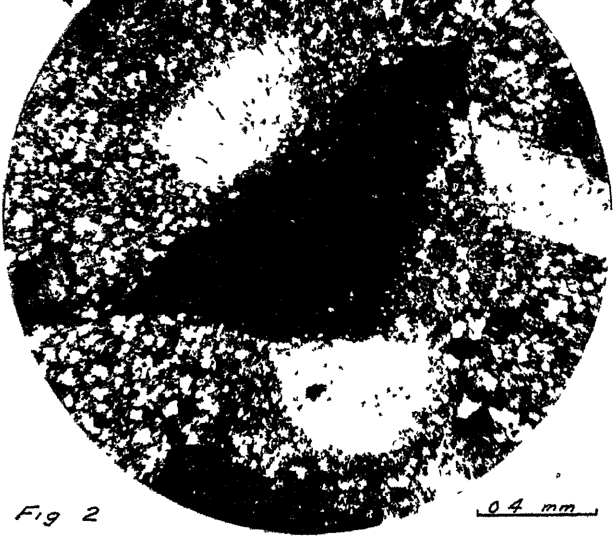


Fig 2

PLATE 2

Fig 1 Incompletely recrystallized sandy red bed argillite Clastic quartz grains in turbid matrix with crystalloblastic epidote, carbonate, and some alkali feldspar Crossed nicols Upper T'aohuats'un valley east of Hsiaoch'iaot'ou, near western contact of "granodiorite-porphyry" (no 28 12 45 4)

Fig 2 Metasomatic "granodiorite-porphyry" Crossed nicols Groundmass of clastic quartz grains cemented by alkali feldspar Early stages of porphyroblastic growth of replacement plagioclase Paiyenssu northeast of T'aohuats'un (no 29 12 43 3)

Fig. 1

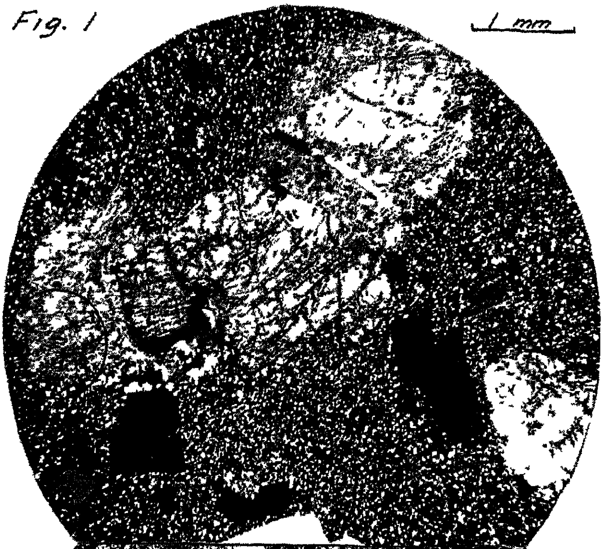


Fig 2

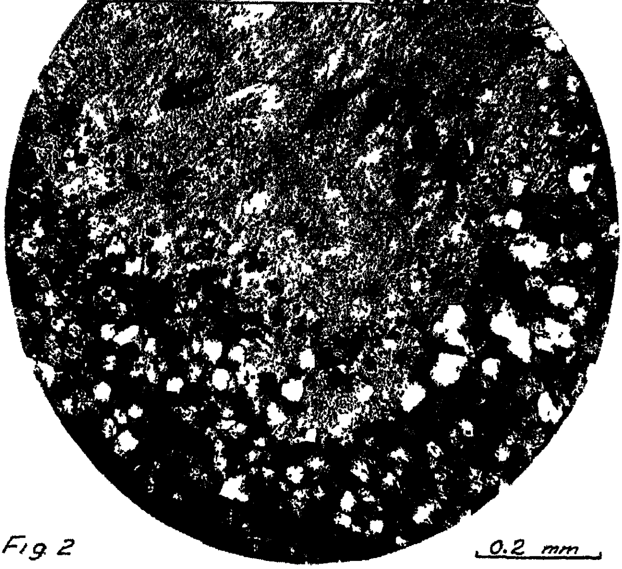


PLATE 3

Metasomatic "granodiorite-porphyry" Crossed nicols Porphyroblastic growth of plagioclase in clastic-derived groundmass East of Chimaoku (no 27 12 15 7)

Fig 1 Large composite plagioclase porphyroblast Below quartz crystal

Fig 2 Detail of the same plagioclase porphyroblast (area marked by circle in fig 1), showing replacement border

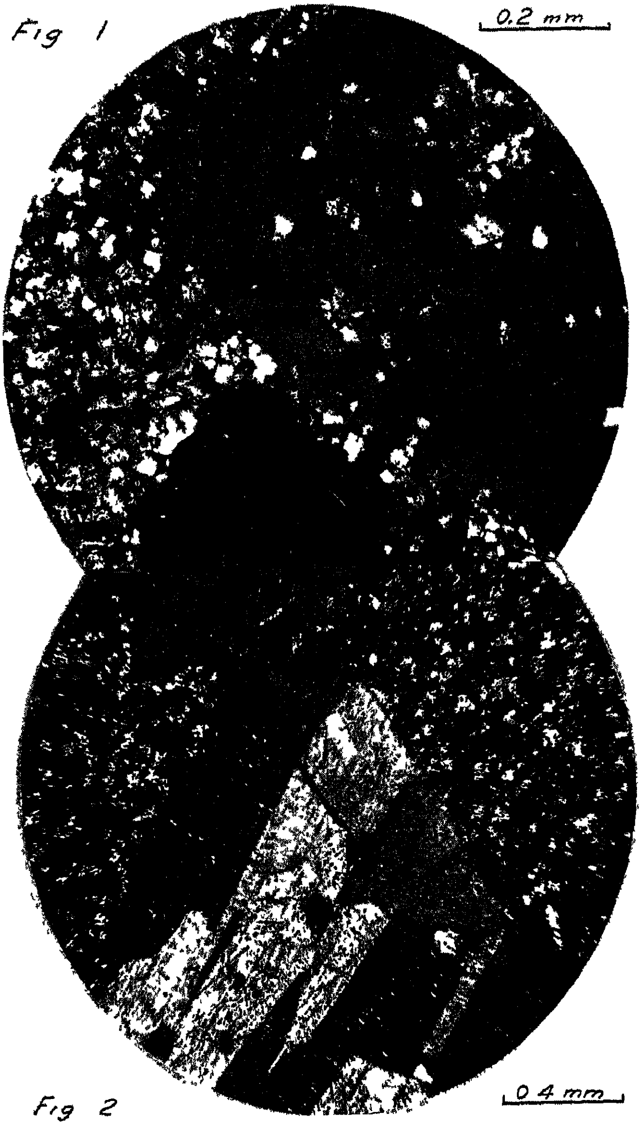


PLATE 1

Advanced stages of porphyroblastic growth of replacement plagioclase in metasomatic "granodiorite-porphyry" Crossed nicols

Fig 1 Crystal faces have developed, but relict inclusions of clastic quartz grains of the groundmass are still present Northwest of Chiench'uan (no 22 1 46 2)

Fig 2 Final stage the plagioclase has acquired crystal form and cleared itself of inclusions, imitates a phenocryst Above a less advanced plagioclase porphyroblast, still showing replacement borders West of T'aohuats'un (no 28 12 45 31)

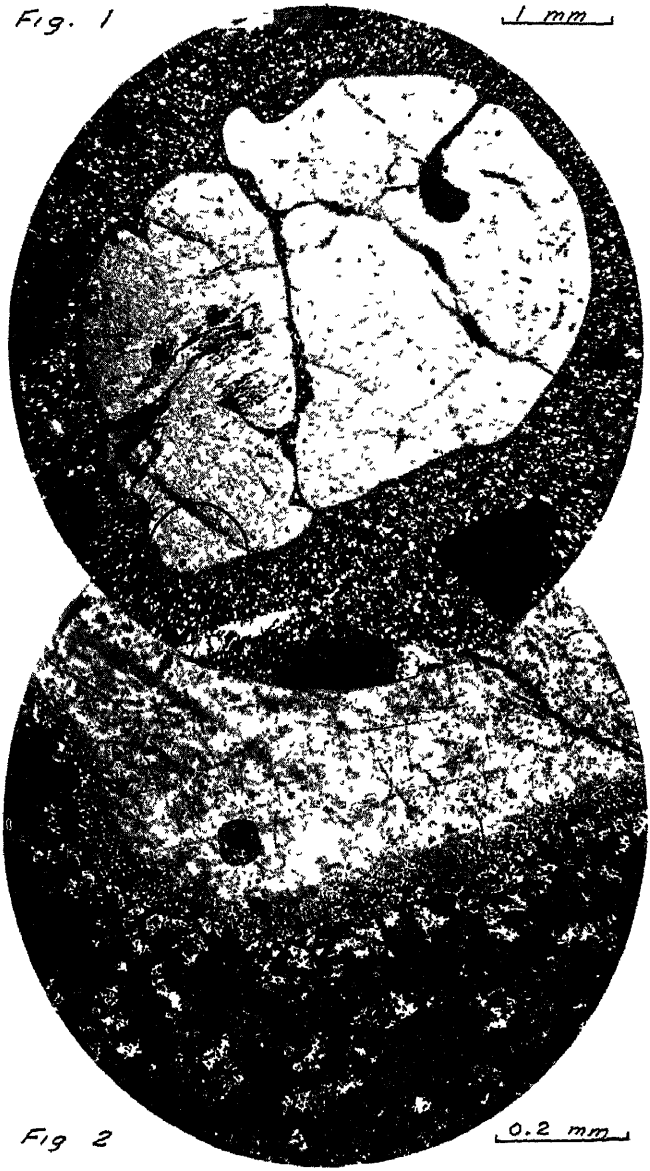


PLATE 5

Advanced stage of growth of quartz porphyroblast in metasomatic "granodiorite-porphyry" Crossed nicols West of T'aohuats'un (No 28 12 45 31)

Fig 1 General view The quartz grain has the appearance of a phenocryst with "resorption embayments"

Fig 2 Detail of the same quartz grain (area marked by circle in fig 1), shown growth surface of porphyroblast invading the groundmass

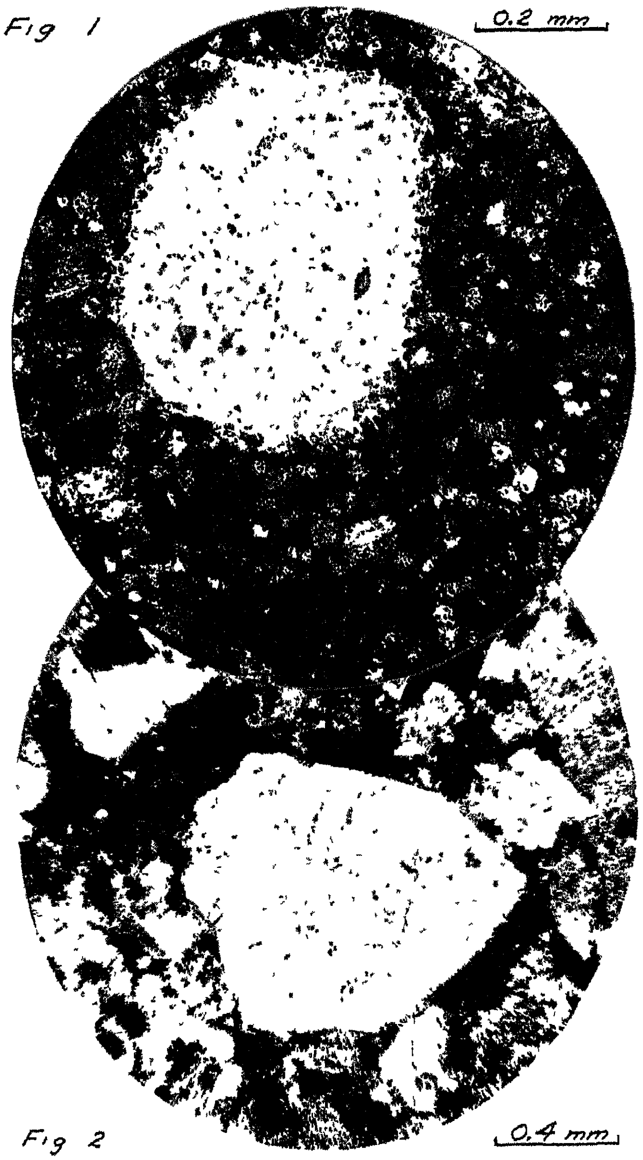
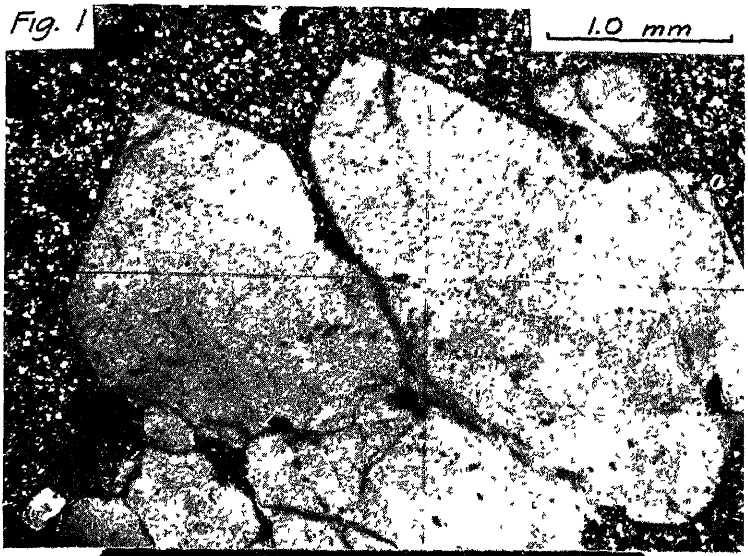


PLATE 6

Fig 1 Early stage of porphyroblastic growth of quartz grain in metasomatic "granodiorite-porphyry" Growth surface as in pl 5, fig 2 Crossed nicols Northwest of Chiench'uan (no 22 1 46 1)

Fig 2 Larger clastic quartz grain before crystalloblastic growth In altered finely breccious argillaceous red bed sandstone Plain polarized light Dark areas are newly formed hornblende South of upper part of T'aohuats'un Valley, south of Hsiaoch'iaot'ou (no 28 12 45 28)

Fig. 1



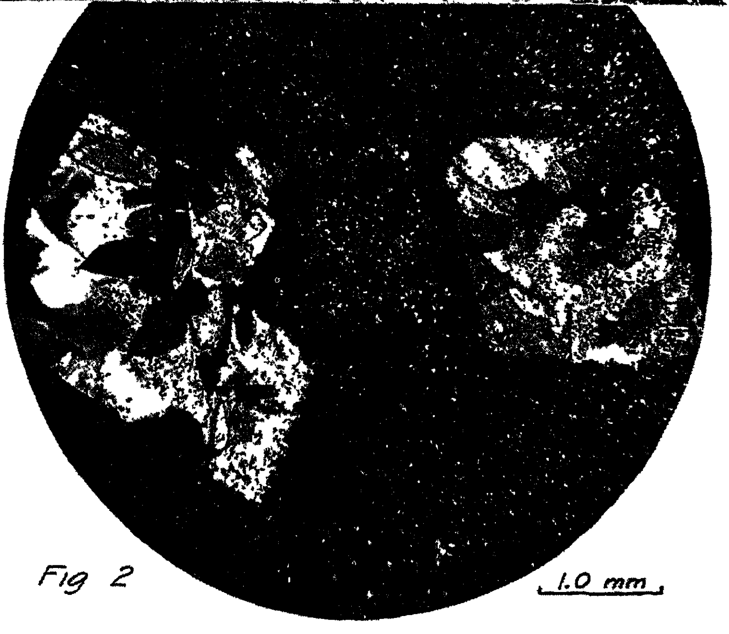


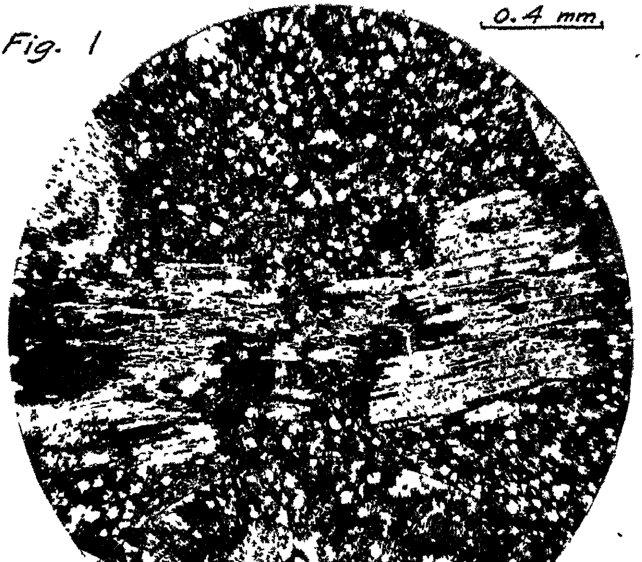
Fig 2

PLATE 7

Fig 1 Advanced stage of porphyroblastic growth of quartz in metasomatic "granodiorite-porphyry" Crossed nicols Between two idioblastic projections of the quartz grain an embayment of relict groundmass has been spared out, imitating a resorption embayment Another projection of the same crystal is xenoblastic and encloses groundmass grains Paiyenssu northeast of T'aohuats'un (no 29 12 45 4)

Fig 2 Incompletely recrystallized very fine-grained clastic red bed sediment, containing "dioritic" crystalloblastic patches of green hornblende and plagioclase Crossed nicols South of upper part of T'aohuats'un Valley, south of Hsiaoch'iaot'ou (no 28 12 45 15)

Fig. 1



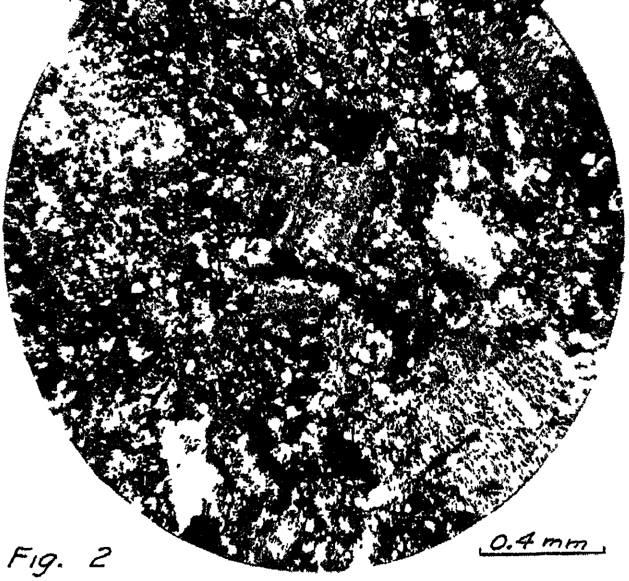


Fig. 2

PLATE 8

Fig 1 Porphyroblastic growth of biotite in clastic-derived groundmass of metasomatic 'granodiorite-porphyry.' Comparatively early stage Crossed nicols East of Chimaoku (no 27 12 15 7)

Fig 2 Incipient stage of sodium metasomatism of Sheku type in fine-grained clastic Triassic sediment south of Mat'oushan (no 18 1 16 5) Crossed nicols Small relict clastic quartz grains, xenoblastic albite, and chlorite

low index of refraction and its low birefringence; for other methods of optic determination, the grain size is too fine). Some quartz grains of the groundmass develop projections of amoebae-like shapes, in the same kind of matrix. Subordinately fine-grained mosaic of recrystallized quartz occurs. In other specimens, also from widely distant localities, the groundmass consists of a felt-like intergrowth of "fluffy" potash-feldspar (determined as above) and quartz, mostly with varying amounts of plagioclase which occasionally has developed into well formed tiny crystals. There are also transitions between the "clastic type" and the "felty type" of groundmass. They show angular clear quartz grains as well as an intergrowth of irregular amoebae-like small crystals and felty aggregates.

Some further textural features will be described in chapter (c) where the replacement origin of these rocks is discussed. The general appearance of most of these rocks is remarkably uniform in the wide area shown in the map (pl. 9). There is no recognizable mineralogical or textural difference between the marginal and the interior parts of the individual outcrops studied. *In a purely descriptive way*, the rock may be named "*granodiorite-porphyry.*"

More basic varieties with little or no quartz were collected locally southwest of Sheku and northwest of this town where they seem to be more common. They contain diopsidic pyroxene which usually shows irregular crystalloblastic growth and pronounced sieve texture though more compact grains occur. Green hornblende of similar texture may replace part of the pyroxene. Subordinate biotite forms small patches in pyroxene and hornblende, and also occurs in irregular grains showing sieve texture. There is much plagioclase, but potash-feldspar may be lacking. The few specimens of this rock do not possess a groundmass comparable to that of the "granodiorite-porphyry" though larger porphyroblasts may be contrasted to an enclosing granoblastic fabric. As stated above, this "dioritic variety" probably belongs together with the more acidic predominant rock type.

(c) *Metasomatic origin of "granodiorite-porphyry"*

A replacement origin of the "granodiorite-porphyry" is suggested by the clastic relict texture of its groundmass and by the porphyroblastic nature of its apparent phenocrysts.

(1) *Clastic derivation of the groundmass*

As the description of the altered quartzose-argillaceous red bed sediments on the one hand, and of the groundmass of the "granodiorite-porphyry" on the other hand suggests, and as a glance at the photomicrographs in plates 1-8 confirms, both show an extremely strong resemblance. Both contain the same small clastic grains of clear quartz. In both rocks, relative amount, shape and size of these grains vary within the same limits. In both rocks, clastic quartz grains frequently show small projections due to beginning growth. In both rocks, these tiny clastic grains are cemented by a more or less turbid and frequently iron-stained matrix. In the altered red bed sediments, this matrix is very incompletely recrystallized, often still containing kaolinitic-appearing material, and sometimes containing low-temperature new minerals such as sericite and chlorite; it is rich in carbonate, and epidote is the chief new mineral whereas alkali-feldspar is subordinate. In the groundmass of the "granodiorite-porphyry," the cement between the clastic quartz grains has on the whole somewhat cleared itself and has mainly recrystallized as alkali-feldspar; carbonate and epidote, as well as most of the magnetite, have disappeared, having been used by the large grains of new hornblende and plagioclase. In altered red beds from the area south of the "type section" (fig. 1), actinolite and hornblende have frequently begun to form before feldspathization set in. The hornblende of these rocks has the same optic properties as that of the "granodiorite-porphyry." Both the altered red bed sediments and the groundmass of the "granodiorite-porphyry" also show a felty or "fluffy" finely recrystallized variety free of, or poor in, clastic quartz grains. In the case of the red beds, the lack of such clastic grains is primary, and the same presumably applies in most cases where the "granodiorite-porphyry" has a similar groundmass.

I conclude that *the groundmass of the "granodiorite-porphyry" is not of igneous origin, but derived from the fine-grained clastic red bed sediments.*

(2) *Porphyroblastic nature of the apparent phenocrysts*

Superficial examination of thin sections of the "granodiorite-porphyry" shows apparent phenocrysts in the fine-grained groundmass. Closer study reveals features which are incompatible with igneous phenocrysts, and suggest porphyroblastic growth in a solid rock. These features are especially well displayed by the *plagioclase.* Plate 3, fig. 1, shows a large composite plagioclase grain, and figure 2 a part of the same plagioclase under higher magnification. The plagioclase invades and absorbs the fine-grained groundmass, and it encloses its small quartz grains as relics. In other words, the groundmass was there before the plagioclase crystal formed. In its interior part, the plagioclase porphyroblast has succeeded in clearing itself of relict inclusions. In all thin sections examined, nearly all of the plagioclase grains exhibit distinct replacement borders of this type. Some plagioclases have preserved relict inclusions of the groundmass even in their interior parts (cf. pl. 4, fig. 1). Early stages of growth of plagioclase porphyroblasts are present in every thin section. The beginning of crystalloblastic growth of a plagioclase replacing the groundmass is shown in plate 1, figure 2. These early crystalloblastic patches are small, are often still turbid, and have ill-defined borders. A somewhat more advanced stage is seen in plate 2, figure 2. Plate 3 illustrates a much more advanced stage. The porphyroblast has become large, but it still lacks crystal form. Finally, crystal faces begin to develop, but relict inclusions of groundmass grains may still be preserved (pl. 4, fig. 1). If idioblastic crystals succeed in clearing themselves of inclusions, they imitate phenocrysts (pl. 4, fig. 2). This is the final stage of feldspathization. It must be stressed that in all thin sections examined which come from marginal as well as interior portions of "granodiorite-porphyry" outcrops, the various stages of development of plagioclase porphyroblasts occur together. Among the specimens collected, not one contains exclusively "good phenocrysts."

Although some incipient porphyroblasts of plagioclase are turbid, larger grains are usually clear, and only occasionally are turbid centers observed. Some plagioclase porphyroblasts are composite, and sometimes several grains are seen to coalesce which usually leads to complex twinning. Many of the large porphyroblasts show complex twinning, and it seems to me that this has to be chiefly attributed to coalescing. In some zoned crystals, coalescing has caused a complex arrangement of the zones. Occasionally, plagioclase porphyroblasts enclose small grains of hornblende, biotite and sphene which had developed prior to feldspathization; it may be recalled that hornblende and occasional biotite also occur in some altered red beds.

The *potash-feldspar* crystals show the same replacement features and stages of development as the plagioclase. Potash-feldspar may form composite porphyroblasts with plagioclase, and occasionally with quartz. More commonly potash-feldspar encloses smaller plagioclase porphyroblasts, indicating that the optimum of potash-feldspar growth was somewhat later than that of plagioclase (cf. below). Apart from small relict grains of groundmass quartz, potash feldspar sometimes encloses small hornblende and titanite crystals.

The large *quartz* grains are also of a porphyroblastic nature although this is much less conspicuous than in case of the feldspar and, in fact, the quartz strongly suggests igneous phenocrysts upon superficial examination. Only exceptionally (pl. 7, fig. 1) do quartzes marginally enclose relict grains of the groundmass. This is understandable if it is realized that the high quartz content of the groundmass made it easy for a growing quartz grain to absorb the groundmass. Almost all of the large quartzes have sharp borders, and they often show those groundmass-filled embayments (cf. pl. 5, fig. 1, and pl. 7, fig. 1) which are so common in the quartzes of porphyritic acidic rocks and which are traditionally attributed to magmatic resorption. However, under higher magnification (pl. 5, fig. 2) the borders of all quartzes, including those having embayments, are seen to be surfaces of growth, with numerous delicate projections growing into the groundmass and absorbing it. Moreover, like the feldspar, the quartz shows all stages of porphyroblastic growth, beginning with small crystalloblastic patches in the groundmass, and ending with bipyramids imitating phenocrysts. A moderately early stage is illustrated by plate 6, fig. 1. This grain exhibits the same type of growth surface as the more advanced large grains. In plate 6, this grain is confronted with a clastic quartz grain of comparable size, occurring in an altered finely breccious red bed sediment. It seems plausible that some of the quartz porphyroblasts in the "granodiorite-porphyry" have started from such clastic grains exceeding the usual small grain size of the sediments. However, this cannot apply to the majority of the quartz porphyroblasts since in the "type section" (fig. 1) most of the red bed sediments laterally prolonging the "granodiorite-porphyry" and enclosed in it, are fine-grained.

The features described suggest that the groundmass-filled embayments are caused not by magmatic resorption, but by *differential growth* of quartz crystals in the solid groundmass. Some portions of a porphyroblast grew faster than others, and relict groundmass is preserved between such crystalloblastic projections. Not all embayed crystals would by their shapes suggest this interpretation (cf. pl. 5, fig. 1; however, some of these quartz grains may be cut close and parallel to their surfaces and therefore not reveal the typical pattern). But other embayed crystals do have shapes suggestive of this interpretation, and occasionally embayed quartzes are so shaped as to constitute definite evidence of growth, in addition to the generally developed surface character indicative of growth (cf. above). Such a case is shown in plate 7, figure 1. Here a quartz porphyroblast has grown differentially, two neighboring crystalloblastic projections have developed crystal faces, and between them an embayment of relict groundmass has been left. It is obvious that magmatic resorption would have attacked the crystal edges. In other words, this is growth, not resorption. This reasoning is borne out by another part of the same quartz individual (pl. 7, fig. 1) which shows an irregular replacement border enclosing relict grains of the groundmass.

I would like to state briefly two general arguments against magmatic resorption of the quartz crystals described. First, it is hard to conceive how a residual magma exceedingly rich in silica from which the quartz-rich groundmass would have to be derived, could have resorbed earlier-formed quartz phenocrysts. Second, it is difficult to understand why magmatic corrosion should form narrow embayments instead of attacking the quartz crystals in a more uniform manner along their surfaces and first rounding off any crystal edges which might have been present. The only possible mechanism of magmatic formation of embayments seems to me to be mechanical cracking of crystals still surrounded by magma (due to a rapid change of physical conditions), flowing of magma into such cracks, and magmatic corrosion along such fractures. However, in the rocks here described, no such systematic relationship between embayments and cracks exists, although later fractures which are widely present, occasionally happen to have followed embayments.

The texture of the subordinate mafics in the "granodiorite-porphyry" matches that of the felsic porphyroblasts. Both *hornblende* and *biotite* show all gradations from xenoblasts to perfect idioblasts. In their early stages, they penetrate irregularly, replace and enclose the fine-grained groundmass. Elongation parallel to the c-axis of the hornblende and the base of the biotite already asserts itself at this early stage. A tendency to develop crystal faces becomes visible quite soon. A comparatively early stage of porphyroblastic growth is illustrated by plate 8, figure 1. Poikiloblastic textures usually still survive when the grains begin to develop continuity and to form well-defined crystal faces. The tendency of the mafic porphyroblasts to develop continuity first in their marginal portions, results in crystals which, in an advanced stage of development, still contain relict grains of groundmass quartz in their central parts. Occasionally this tendency has led to the formation of idioblastic crystals with a hollow core filled by groundmass fabric.[6] In the final stage of porphyroblastic development, the mafic crystals completely clear themselves of relict inclusions. The result is perfectly euhedral and continuous hornblende and biotite crystals indistinguishable from true igneous phenocrysts. It must be emphasized that, as in the case of the felsic porphyroblasts, early and advanced stages of development of mafic porphyroblasts occur together in every specimen examined.

As supplementary evidence, a rock from a neighboring area may be briefly described which shows an *incipient stage of granitization of Sheku type*. It was collected about 8 miles southeast of the southeastern corner of the area shown in the map (pl. 9), in a mountain range extending south from a peak called Mat'oushan. Here Upper Triassic (Norian) marine shales and sandstones contain some smaller sill- or dike-like bodies of light-colored igneous-appearing rocks which resemble fine-grained diorite-aplites. The microscope shows several varieties most of which bear clear evidence of a metasomatic origin. The rock here described is illustrated in plate 8, figure 2. It consists of a typically clastic fine-grained matrix and a large amount of albite and a smaller amount of chlorite porphyroblasts. The groundmass consists of angular and subrounded clear quartz grains part of which show small growth projections. They are cemented by an interstitial mass chiefly composed of albite. Most of this cement belongs to small incipient albite porphyroblasts each of which surrounds a number of clastic quartz grains.

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[6] Somewhat similar growth has been described (in plagioclase porphyroblasts) by G. E Goodspeed (1937b, p 1135)

The albite (An 3) porphyroblasts show pronounced sieve texture, enclosing clastic grains of groundmass quartz as well as smaller grains of crystalloblastic chlorite. The albite grains have irregular outlines. There are incipient as well as more advanced stages of porphyroblastic growth, but the stage has not yet been reached at which crystal faces begin to form, nor do the albite grains attain the size which is shown by the large plagioclases in the rocks from the Sheku region. Many of the albite porphyroblasts have turbid centers often containing sericite, and clear margins. Others are altogether turbid. The chlorite occurs in xenoblastic grains forming irregularly shaped aggregates with crenulated borders which marginally penetrate the groundmass, as well as in individual porphyroblasts. The latter vary in size and are mostly xenoblastic, with crenulated borders, but some of the larger grains show idioblastic tendencies and develop a platy habit. Part of the chlorite is associated with sericite which is often concentrated in the interior of chlorite aggregates. The rock also contains amoebae-like grains of calcite. Some are small, others become porphyroblastic. Concentrations of calcite and chlorite occur together, indicating a larger admixture of dolomite in certain portions of the clastic sediment. Another specimen shows micrographic textures formed by metasomatic replacement.

The original sediment was a very fine-grained highly argillaceous sandstone or siltstone, containing some carbonate. The chlorite is undoubtedly derived from the constituents of the original sediment. The large quantity of albite, on the other hand, cannot be of the same origin, because no normal argillaceous-sandy sediment has such a high sodium content. Sodium must have been introduced. It has reacted with the argillaceous material in the sediment. The result was albitization. Both the presence of chlorite as a mafic, and the association albite plus calcite (without formation of calcium-bearing plagioclase), indicate that feldspathization has in this case taken place under very low temperature, corresponding to epizonal temperatures in regionally metamorphic rocks. We may speak of local low-temperature sodium-metasomatism in a fine-grained clastic sediment—an incipient stage of feldspathization of the type developed on a large scale in the Sheku region.

IV. GENETIC DISCUSSION OF GRANITIZATION IN THE SHEKU REGION, AND COMPARISON WITH OTHER TYPES OF STATIC GRANITIZATION

(a) *Summary of unusual features in the Sheku region*

The granitized bodies described show several remarkable features. (1) Owing to the practically undisturbed condition of the enclosing sediments, the process of static granitization is displayed in an unusually convincing manner. (2) This mode of origin can be demonstrated for the whole of the granitic masses and not only their marginal parts. (3) Metasomatic transformation has here taken place at a remarkably shallow depth. (4) The non-granitized sediments enclosing the granitized bodies do not show thermal metamorphism of regional extent as is common in other areas of static granitization, and recrystallization of the sediments at contacts is surprisingly restricted. (5) A sediment-derived groundmass showing clastic relict texture has in a rather uniform manner survived in all parts of the granitized bodies. (6) The granitized rocks, as a whole, show a surprisingly uniform character, and contacts are usually sharp.

Some of these items have been sufficiently discussed in chapters II and III, others will have to be elaborated below. Points (4) and (6) are rather unexpected in metasomatic granitization and could indeed be used as arguments against it, if the evidence of granitization in the Sheku region were less convincing. I shall try to show that these features, and also the features listed as (5), are due to special conditions in this region, essentially the shallow depth of granitization, combined with the nature of the original sediments and the lack of deformation.

(b) *Shallow depth of granitization*

The granitized bodies of the Sheku region occur in the "upper red beds" (cf. above). These are the latest Mesozoic sediments I have observed anywhere in Northwest Yunnan. The scantily developed Tertiary formations are later than the granitized rocks here described (cf. chapter I).[7] In view of the intense folding which the "upper red beds" have suffered in wide parts of Northwest Yunnan, it is extremely unlikely that any Mesozoic formation younger than the "upper red beds" and older than their folding and their granitization, could have existed without having in some areas been preserved in synclinal cores.

Therefore, the depth at which granitization took place at Sheku can hardly have been greater than the thickness of that portion of the "upper red beds" which has been eroded above the granitized bodies. The red beds preserved west of Sheku are about two miles thick. The granitized rocks exposed near the floor of the T'aohuats'un valley (traverse shown in fig. 1, southern section) were formed about one mile below the highest preserved red bed member, but farther north (fig. 1, middle section) similar granitized rocks represent the highest preserved red bed member. The question therefore is: How thick was that part of the "upper red beds" which has been entirely eroded? No direct evidence is available. However, it seems reasonable that it could hardly have exceeded a mile or two, in addition to the two miles of "upper red beds" preserved. This estimate is borne out by some sections of "upper red beds" which I have observed in other parts of Northwest Yunnan, where the thickness does not exceed a few miles.

The depth of granitization at Sheku was approximately between one and three miles, if the present vertical range of exposure of more than one mile is taken into account, in addition to the estimated thickness of one to two miles of the entirely eroded red beds. For granitization on a regional scale, i.e. for the *making* of granitic rock, this is a shallow depth indeed. In most cases, regional granitization has acted at greater depths. Obviously, the depth at which granitic rock is *made* must not be confused with the depth of granitic intrusion, i.e. of the secondary emplacement of ready-made granitic rock which has acquired mobility and has, by mechanical means, risen from its place of origin (cf. the brief discussion on "rheomorphism" at the end of Part III).

[7] In the southeastern part of the area shown in pl. 9, occurs a volcanic formation which is probably Tertiary though it might be Cretaceous It overlies moderately folded "upper red beds" with angular unconformity, and is certainly not older than the granitized rocks of Sheku.

(c) *Mineralogical-chemical balance of granitization*

A comparison of the mineral composition of the incompletely recrystallized red beds and of the "granodiorite-porphyry" at Sheku shows that here the principal chemical feature of granitization was *addition of sodium.* The subordinate amounts of hornblende and biotite are derived from the original sediment. Iron was supplied by the hematite which is responsible for the color of the red beds and has been made into magnetite in the altered red beds Calcium and magnesium occur as carbonate admixed in the clastic sediment. Carbonate is also present in most samples of altered red beds, though part of the calcium has here gone into epidote, and in some specimens from south of the T'aohuats'un valley some (exceptionally all) calcium and magnesium have been absorbed by amphiboles. In the granitized rocks, all of the magnesium and most of the iron—except a smaller amount persisting as magnetite—have entered hornblende and biotite. Some of the calcium has been consumed by hornblende, but most of it has been used for the anorthite component in the plagioclase. Neither free carbonate nor epidote have survived in the granitized rocks. This is not surprising, in view of the very large amount of metasomatic plagioclase formed. A minor quantity of calcium was used in sphene to bind titanium.

Silica and aluminum are the main constituents of the quartzose-argillaceous sediment. A minor amount of aluminum was used for the subordinate biotite of the granitized rocks, and the hornblende also consumed a small quantity of aluminum. Most of the aluminum went into feldspar. Silica was available in excess. Where, as for instance in the hornblende and the oligoclase, more silica was required relative to aluminum than was present in the clay minerals of the sediment, it was provided by the clastic quartz grains. Most of these have been absorbed in areas occupied by mafic as well as feldspar porphyroblasts. Their silica has, in part, been used for the porphyroblasts, and in part been pushed out to contribute to the porphyroblastic growth of quartz. The minor amount of potash needed for the subordinate biotite was supplied by the argillaceous material in the sediment (adsorption of potash). This is confirmed by the occurrence of sericite and occasional biotite in incompletely recrystallized red beds. That hornblende and biotite of the granitized rocks are derived from the original sediment, is borne out by the presence of these minerals in some of the altered red beds from the mountains south of the main traverse surveyed. In these rocks, actinolitic amphibole has formed prior to hornblende. It represents a stage when temperature was lower.

The granitized rocks contain large amounts of porphyroblastic plagioclase. The calcium needed for the anorthite component was provided by the original sediment (cf. above). The sodium required for the large amount of the albite molecule present, could not be supplied by the original rock, since no normal quartzose-argillaceous sediment is nearly that rich in sodium. Addition of sodium is evident. Apart from sodium, probably a smaller amount of potash was introduced, though part of the potash entering potash-feldspar was provided by the sediment (the very minor amount of biotite would not utilize all the potash present in the argillaceous sediment). In addition to the potash-feldspar in the groundmass cement, the large microperthite porphyroblasts strongly suggest addition of some potash. It would have been introduced at a late stage of granitization; for it has been shown above that the optimum of porphyroblastic growth of potash-feldspar was later than that of plagioclase, though there is a wide range of overlap in the time of formation of these two minerals.

Muscovite is absent in the granitized rocks, although the original sediment is rich in clay and often contains sericite where incompletely recrystallized. Biotite occurs only in minor amounts in the granitized rocks, most of the iron and magnesium being utilized in hornblende which requires less aluminum. Nearly all the aluminum of the original sediment has been used for the formation of feldspar. It is thus indicated that the originally alkali-deficient rock became, during granitization, saturated with alkali. In other words, the granitizing agent was characterized by an alkali excess, with sodium predominating. As the mineralogical-chemical balance shows, the plagioclase, and even the albite molecule, were not introduced as a whole, but essentially sodium was the only substance added in the process of plagioclase formation. The plagioclase porphyroblasts are the product of reaction between country rock and granitizing agent, and are therefore truly metasomatic. This is confirmed by their replacement texture.

(d) *Mechanism of introduction*

For synkinematic granitization, this question has been discussed in Part I of the present paper (p. 226). Fundamentally, the mechanism is similar in static replacement. The main difference is that the directional influence of active differential shearing on the movement of "solutions"[8] is absent in static metasomatism, i.e. in infiltration of a mechanically isotropic rock mass. Therefore, the granitizing fluids tend to spread out in a more irregular manner. Structures which are mechanically dead but are still present, may have a certain directional influence during static granitization. They are either primary, such as bedding or different composition of adjacent strata. Or they are secondary but pre-metasomatic, such as schistosity of rocks which were metamorphosed prior to their static granitization, or as fractures, crush zones and joints, the latter three having only a local directional influence. In the Sheku region, the irregular spread of granitization is very pronounced, because the infiltrated sediments are comparatively uniform in character, and have not acquired secondary anisotropy by intense deformation prior to granitization.

Another difference in static granitization is that there is no mechanical breakage of mineral grains which under conditions of synkinematic metamorphism and replacement facilitates the movement of solutions. On the other hand, under static conditions, the lack of tectonic compression would tend to make it easier for solutions to rise.

In Part I, certain general arguments were advanced in favor of transportation in "solution" of the introduced material, and it was suggested that dry diffusion is not the principal mechanism of long-range transfer although there are examples of its local action in crystal lattices. These argu-

[8] The term "solution" is here, as well as in Part I, used to indicate a mobile phase capable of moving through solid rocks It is supposed to contain both volatiles, essentially water, and metals such as alkalies and silica No implication as to the physical state of these mobile agents is intended. In case of metamorphism and metasomatism under lower temperatures, there will be "ordinary" aqueous solutions. At high temperatures applied under considerable pressures—particularly if temperature rises above the critical point—the behaviour of "solutions" will be modified, and it must also be taken into account that they do not occur as uniform bodies but infiltrate the country rock in an excessively thin intergranular film. It is reasonable to assume, and is supported by ample petrographic evidence, that the hotter these "solutions" are, the more active they become

ments apply equally in the case of static granitization here described. The presence of at least some water during the process of granitization is also here demonstrated by the occurrence of hydrous minerals in the granitized rocks, apart from the fact that the original argillaceous sediments contained considerable amounts of water which were set free in the process of metasomatic feldspathization. It appears unlikely that rocks containing water could have been a suitable medium for dry diffusion as the sole or principal mechanism of long-range transfer.

Generally speaking, it seems to me that the concept of an active liquid phase, moving in an intergranular film, is supported by the microscopic study of both synkinematically and statically granitized rocks, especially if rocks showing incipient stages of granitization are studied. Most sections cut from such rocks clearly demonstrate intergranular activity including transfer of material; resorption of pre-existing crystals; reactions between original and introduced material; and the formation of new crystals. Dry diffusion as the principal mechanism would not be concentrated at the crystal boundaries, i.e. in the intergranular, but pass through the crystal lattices. As far as dry diffusion plays a role, it appears easier to conceive of it as a subsidiary mechanism which acts on a local scale in single crystals and small groups of crystals. Instances where microscopic textures suggest dry diffusion on the scale of single crystals, are numerous. An example given in Part I are statically recrystallized norites from the Nanga Parbat area in which crystalloblastic hornblende has penetrated calcic plagioclase, and this latter has near the hornblende been albitized, the plagioclase lattice remaining intact. The same process was observed in these rocks where crystalloblastic epidote penetrates calcic plagioclase. Among textures described from the granitized rocks of Sheku, the "pushing out" of surplus quartz from metasomatic plagioclase and other porphyroblasts suggests dry diffusion on a local scale. What has been said above about the mechanism of transfer, applies to shallow and moderately deep-seated metasomatism only; for the rocks used as evidence have formed at such depths (cf. also Part I). At very great depths, dry diffusion could more easily be the principal mechanism of transfer (cf. J. A. W. Bugge, 1945).

The location of the Sheku area of granitization seems to be determined by the regional structural pattern. As far as I have mapped Northwest Yunnan, much of this region is characterized by a northward divergence of orogenic trends, part of which is shown within the map (pl. 9). On the east, the trend is a little east of north, and on the west it is north-northwest. In the angle between these diverging compressive structures the "upper red beds" lack any strong deformation, but farther south they become involved in intense deformation and dynamic metamorphism, where the eastern and western structural elements finally converge. Because of this divergent pattern, the area of Sheku has not been compressed after deposition of the "upper red beds," and presumably this was the reason why here granitizing solutions rose more easily to high levels from the depths of the geosyncline, whereas they did not penetrate the highly compressed "upper red beds" farther south.

(e) *Substances transferred, and volume question*

As to the substances present in the granitizing solutions, I again refer to the discussion in Part I (p. 239-241). The general points made there also apply in the case of static granitization here described. The chief difference is that at Nanga Parbat the predominant alkali introduced was potash, whereas at Sheku it was sodium. However, both synkinematic and static granitization comprise examples of potash as well as of sodium predominance, and some intermediate cases. The differential character of granitization controlled by the composition of the country rocks, which has been demonstrated at Nanga Parbat, is not well illustrated in the statically transformed red beds of Sheku because they are comparatively uniform in composition. However, where static granitization has acted on rocks of highly varied composition, the principle of differential granitization is often as clearly displayed as at Nanga Parbat. An example is the area between Riverside and Conconully in North Central Washington (cf. Part III, chapter III).

As at Nanga Parbat, it can be demonstrated at Sheku that *the material added to the country rock did not have the composition of granite or granodiorite, nor even of feldspar. The essential material added was alkali.* In the metamorphic argillites of Nanga Parbat, *varying amounts of silica* were precipitated with the alkali. To what extent silica has been added at Sheku, is not easy to ascertain because the red beds affected contain variable, usually comparatively large, amounts of clastic quartz. As far as silica has been precipitated, its amounts have been small. Silica—presumably in large part derived from quartz-rich members among the geosynclinal sediments—seems always to accompany the alkali in granitizing solutions. For in areas of granitization rocks which compared to granite or granodiorite were deficient in silica, have generally been able to obtain the silica required. No silica is precipitated in rocks high in silica, and the amount precipitated in other rocks depends on their needs, i.e. on their silica deficiency compared to granite or granodiorite.

The amount of alkali added in the granitized rocks of Sheku was several per cent. Considering the primary quartz content of the red beds, the *total amount of added material* hardly exceeded about five per cent. Equally small amounts have been added during the granitization of other quartzose-argillaceous sediments, if their original silica content did not differ much from that of a granite or granodiorite.[9] Argillites poorer in clastic quartz, i.e. relatively purer shales composed of clay minerals with an admixture of quartzose silt, and their metamorphic equivalents, on the other hand, contain less silica than granite or granodiorite. Many of these rocks need as much as 10 to 15 per cent of silica to become granitized. Thus, in rocks of the series argillite—quartz-rich argillite, the *amount of silica added during granitization varies* from about 15 per cent to zero. Sediments richer in silica than granodiorite or granite will during granitization undergo a silica decrease, either relatively, or absolutely, i.e. due to actual removal of silica. Obviously certain arkoses do not require any addition at all during their granitization.

As at Nanga Parbat, there is no sign in the Sheku region of substances having been *removed* from the metasomatized rocks (except water, cf. below) and precipitated in an outer aureole surrounding the granitized bodies. Generally speaking, it appears to me that *large-scale outer fronts* formed by precipitation of substances driven out from the country rock, are *frequently lacking* both in areas of synkinematic and of static granitization, and that therefore the formation of a "basic front" cannot be considered as a *necessary* part or result of granitization. That this process *can* happen, has been demonstrated by Doris L. Reynolds (1946; 1947). In my personal field experience, I have observed *local* concentrations or segregations of mafic constituents which, compared to the size of the granitized bodies, are small—as a rule, extremely small—and could be described under the heading "local metamorphic differentiation."

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[9] Quartzose-argillaceous sediments containing about 45 to 50 per cent quartz will require neither much introduction nor much removal of silica during their granitization

In all granitized areas where addition is not compensated by removal of material, the *volume* of the transformed rock body obviously must have increased. However, as far as *granitization of argillaceous rocks* is concerned, increase of volume due to addition of material must be supposed to be to some extent compensated by loss of water which is set free during the metasomatic formation of feldspar from hydrous minerals such as clay minerals and micas. If a calculation is based on a mineral like kaolinite, the loss of water is considerable and, in per cent of weight, of the same order as the amount of alkali introduced. At the same time, the densities of clay minerals and alkaline feldspars do not differ much. It may therefore be supposed that in those quartz-rich argillites the original quartz content of which made addition of any larger amounts of silica during granitization unnecessary (cf. above), the increase of volume was negligible.

On the other hand, a calculation may be based on metamorphic hydrous minerals such as muscovite, i.e. the feldspathization of rocks may be considered which prior to granitization had undergone a thorough metamorphic recrystallization—a process which was accompanied by a decrease in volume implied in the alteration of sedimentary clay minerals like kaolinite to metamorphic minerals like mica (loss of a considerable part of the water, and somewhat higher density of the mica).[10] On this basis, i.e. considering formation of feldspar from a mineral like muscovite, the amount of water set free is much smaller, and is to some extent counteracted by the somewhat lower density of alkaline feldspars compared to that of micas. Therefore, the volume must have somewhat increased. It must not be overlooked that formation of alkaline feldspar from either clay minerals or mica implies consumption of additional silica. In quartzose-argillaceous rocks with a sufficiently high original quartz content, this silica is furnished by quartz grains which are consumed in the feldspar reaction. As has been shown above, the amount of additional silica required in granitization of rocks of the *series argillite—quartz-rich argillite*, varies from about 15 per cent to zero, depending on the amount of clastic quartz in the sediment, whereas alkali is introduced in all cases.

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[10] Such a slight decrease in volume will be difficult to ascertain, especially where the rock has at the same time been deformed, i.e. in phyllite or micaschist.

In areas where no removal of material other than water during granitization can be demonstrated, therefore only a negligible to moderately small increase of volume has to be expected in rocks of this series. The variation essentially depends (1) on the amount of silica which had to be introduced during granitization, and (2) on the kind of the aluminous minerals used for the making of metasomatic alkaline feldspar, i.e. whether an argillite was more or less directly metasomatized and metamorphic hydrous silicates, formed from clay minerals, only occurred as a transitional phase localized at the advancing front of granitization, or whether a previously metamorphosed argillite was granitized. In the quartzose-argillaceous sediments of Sheku, no large amounts of additional silica were needed during granitization (cf. above), and there had been no regional metamorphic recrystallization prior to granitization. Therefore, the increase in volume, if any, must have been extremely small.

In all those granitized rocks in which a noticeable increase in volume must be expected, such an increase will be more easily ascertained in an area of static than in one of synkinematic granitization. In the latter case, superimposed tectonic deformation tends to obscure structures which might be indicative of an increase of volume. Yet some features indicating increase of volume may be recognizable in areas of synkinematic granitization. In Part I, the structures of certain metasomatic gneisses at Nanga Parbat, such as augengneisses, have been interpreted in this way (p. 227). In addition, the regional structural pattern of the Nanga Parbat gneiss massif as shown in the map (fig. 1, Part I) suggests a certain increase of volume.

Among rocks commonly granitized in geosynclinal belts, the argillaceous sediments just discussed range first. Second are *moderately basic and intermediate rocks* of either sedimentary or igneous derivation. The amount of material added during their granitization is larger, but in most cases ranges from about 25 to about 50 per cent. Rocks of this group are mostly granitized in areas of predominant sodium metasomatism. The products are therefore more often granodioritic and quartz-dioritic than granitic *sensu stricto.* The basic process is metasomatic formation of sodic plagioclase; the material used for it is in part derived from the country rock. Silica is also added. Its continued introduction sooner or later leads to precipitation of quartz. Where potash-feldspar is formed, usually it appears only at a comparatively late stage, and only sometimes in such quantities that the rock becomes granitic *sensu stricto.* As mentioned in Part I, potash metasomatism immediately applied to basic rocks, achieves only biotitization in the majority of cases. As to the volume balance in the granitization of moderately basic and intermediate rocks, either considerable quantities of material must be removed; a process which will be traceable in the field (I again refer to Doris L. Reynolds' "basic fronts"). Or the volume of the granitized body must be considerably enlarged; structural evidence of which detailed field work can hardly fail to produce, especially if transformation has been static.

Granitization of *highly basic and ultrabasic rocks* is rare but does occur. Granitized bodies of this derivation are usually small. Where I have seen transformation of rocks of this group, frequently gabbroid and dioritic stages intervene, but there also seem to be cases of direct introduction of alkali feldspar. In granitized rocks of this group, the added material very commonly exceeds the original material. Removal of part of the original material will be the rule.

(f) *Temperature of granitization, mechanism of heat introduction, and "contact action"*

The granitized rocks of Sheku were formed at a fairly high temperature, corresponding approximately to temperatures in the middle zone of regional metamorphism, and presumably its hotter part. This seems indicated by the mineral association of calcic oligoclase, microperthite, hornblende and biotite. As far as my field observations go, temperature nowhere seems to have become high enough to permit the granitized material to flow. In connection with the temperature of metasomatism at Sheku, it is of interest to recall that southeast of this area local sodium metasomatism has taken place at a much lower temperature, corresponding to that in the epizone of regional metamorphism (cf. p. 390-392). There are other examples of low-temperature feldspathization recorded in the literature. Among them the metasomatism of a coaly sediment described by G. E. Goodspeed, R. E. Fuller and H. A. Coombs (1941) from near Mount Rainier is of particular interest. All cases of feldspathization at very low temperatures seem to be local. Regional granitization appears always to be connected with higher (at least "mesozonal") temperatures, as the mineral assemblages suggest.

The heat required for granitization was transported by the granitizing fluids. What in Part I (p. 241-244) has been said about this point in case of synkinematic granitization, also applies here. In the Sheku region, the identity of the chemical and the thermal agent of granitization is even more obvious. For wherever the sediment has been supplied with the chemical substances required for granitization, simultaneously enough heat has been introduced to allow the characteristic mineral assemblage to form. On the other hand, there has been no perceptible regional increase of temperature in the rock masses not reached by the granitizing agent. Even in the immediate vicinity of granitized bodies, temperature fell off rapidly. Along contacts, the red beds have only been incompletely recrystallized and show, for the most part, a low-temperature mineral assemblage. Moreover, the zone of altered red beds is in most places narrow and sometimes altogether absent. The *rapid decrease of temperature at the contacts of the granitized bodies* is surprising and, in comparison with most granitized areas, unusual. It appears to me to be due to the remarkably shallow position of the granitized bodies in this region (cf. above, chapter IVb), as a result of which heat distributed by conduction was lost very rapidly. Only where hot granitizing solutions spread laterally, would heat not have so easily escaped from beneath metasomatized bodies thus formed, and wider "contact zones" could thus have originated (as probably at the peak 17,000′ in pl. 9).

Most areas of static granitization show evidence of *regional increase of temperature* in the country rocks which have thus been more or less completely recrystallized (hornfelses and related rocks, showing high-grade and more often medium-grade mineral assemblages), if they had not been altered to crystalline schists during preceding orogeny. The distribution of temperature in the Sheku region illustrates that regional heating cannot be considered as primary and static granitization as an ultimate result, but that the reverse relationship exists. This argument does not apply at very great depths where no introduction of heat is required to reach high temperatures. At very shallow depths, very little regional heating may be achieved outside the actually metasomatized rocks. At Sheku, the metasomatizing fluids must initially have been rather hot, and their upward migration must, in terms of geologic time, have been rapid, and there must for some time have been a continuous supply of these hot fluids. Otherwise the great differences in temperature deduced above could neither have been produced, nor have been maintained for some time. Perhaps additional heat was created by chemical reactions after the process had been started by the granitizing agent.

(g) *Preservation of relict textures, and porphyritic-appearing versus eugranitic-appearing textures*

In the groundmass of the granitized rocks of Sheku, a sedimentary relict texture has been preserved. Survival of pre-granitization textures and mineral grains is common in granitized bodies, and usually much more so in their marginal than in their interior portions. Pre-granitization textures more often survive in statically transformed rocks, whereas they are more likely to be destroyed by differential deformation in synkinematically granitized rocks. In this respect, the survival of relict textures in the rocks of Sheku is not surprising. What is unusual is that these relict textures are clastic, and that they have survived throughout the transformed rock bodies.

The preservation of clastic textures is obviously due to the lack of regional heating and consequently of regional thermal metamorphism of the red bed sediments outside the bodies actually reached by the granitizing agent. This feature has been treated above, and has been attributed to the unusually shallow depth of granitization at Sheku. As a rule, in areas of large-scale static granitization, sediments are more or less thoroughly recrystallized as hornfelses and related rocks, and it is therefore the textures of these rocks which are preserved as relics in the granitized rocks, and not the original sedimentary textures which seldom survive in thoroughly recrystallized rocks, although sedimentary structures such as bedding, lamination, etc., are widely preserved in the recrystallized rocks and can sometimes be traced into the granitized rocks. Where static granitization has affected rocks which had previously been subjected to orogenic regional metamorphism and thus become crystalline schists, their schistosity often survives in the granitized rocks as a relict structure which gradually becomes fainter and fainter and may finally disappear (J. J. Sederholm's "nebulites"). At Sheku, the red bed sediments were neither regionally hornsfelsic nor schistose, and therefore their original sedimentary texture was inherited by the statically granitized rocks.

The preservation of a relict groundmass between newly formed porphyroblasts, is responsible for the *apparent porphyritic texture* of the granitized rocks of Sheku. Similar textures often occur in granitized rocks. If the process was synkinematic, such textures are seen in porphyroblast-schists and augengneisses. If replacement was static, textures of a more typically porphyritic appearance are displayed by feldspathized hornfelses, etc., and sometimes by feldspathized rocks still possessing sedimentary relict textures. However, in the majority of cases, such porphyritic-appearing rocks occur in the marginal parts of statically granitized bodies, and larger bodies usually develop less uneven-granular and more *"eugranitic"-appearing textures*, especially in their interior parts. The lack of such eugranitic-appearing textures in the metasomatized red beds of Sheku is remarkable. I would attribute it both to the shallow depth, at which the granitized rock became rapidly cooled after the metasomatizing solutions ceased to rise from below, and to the originally fine-grained and subsequently poorly recrystallized nature of the sediment subjected to metasomatism; the low degree of recrystallization was also due to shallow depth. It seems obvious that in a less fine-grained sediment, and particularly in a thoroughly recrystallized rock like a hornfels, feldspathization combined with more coarse-grained recrystallization of non-metasomatized minerals, will more easily achieve a eugranitic-appearing texture, than in a rock composed of very small clastic quartz grains cemented by altered argillite material.

Generally speaking, we may contrast *transformation of the Sheku type* with a *"normal type" of static granitization.* The type of Sheku is less complete—not in a chemical but in a textural sense—and lacks regional heating of the non-metasomatized country rocks. Both features are chiefly due to the shallow depth of formation. The "normal type" produces more evenly medium- or coarse-grained granitic and granodioritic rocks, and regional heating of the non-metasomatized country rocks is in most cases evident.

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(*To Be Continued*)

SPHEROIDAL WEATHERING OF IGNEOUS ROCKS

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ABSTRACT. Spheroidal weathering is that process by which igneous rocks scale off in concentric shells producing rounded boulder-like forms. Despite recent observations and experimental data, many individuals still misunderstand the true nature of this process. The present paper is an attempt to clarify the matter.

The first part of the paper reviews the several hypotheses which have been advanced from time to time to explain spheroidal weathering. The current views, held by the best authorities, are that spheroidal weathering results from either of two processes: (1) the swelling and spalling of the exterior surface of angular blocks of rock due to the hydration, carbonation, and oxidation of constituent minerals in the outer surface, or (2) the swelling and spalling of the exterior surface of angular blocks which have been subjected to sudden, intense heating of their exterior by forest fires and brush fires

The second part of the paper describes the mineral alterations observed in the shells surrounding several spheroidally weathered boulders. This study shows that in general the core of each boulder consists of relatively fresh rock, but in the shells surrounding the core the primary minerals are altered by weathering to secondary minerals such as kaolinite, sericite, montmorillonite, serpentine, chlorite, hematite, and limonite As a result of these changes the altered rock has increased in volume, the rock in the outer shells having suffered a greater increase than that in the inner ones It is concluded that this increase has caused the rock to scale off in concentric shells

INTRODUCTION

IT is a commonly observed fact that many igneous rocks tend to scale off along joint faces, especially at corners and edges formed by intersecting joints, and thereby develop rounded boulder-like forms which are often but not always enclosed in concentric shells. These rounded boulders range in diameter from a few inches to several feet, and the shells that bound them are from a fraction of an inch to several inches thick. The outer shells normally show a greater degree of weathering than the inner ones, and the core generally consists of fresh, unaltered rock. These boulders have been observed both at the surface and below the surface down to a depth of approximately 40 feet. Examples of this type of weathering have been reported from such widely scattered localities as California, Montana, Vermont, Oklahoma, Hawaii, Ireland, India, and Southern Rhodesia.

The term "exfoliation" has been applied by various writers to this phenomenon, a word which can be traced to the Late Latin *exfoliare* which means to strip of leaves or to scale or flake off. This term has lost much of its significance, however, since it has been applied in a general way to several types of rock spalling, such as the flat flaking of the pyramids in Egypt (Barton, 1916); to sheeting which is flaking on a much grander scale (Gilbert, 1904; Miller, 1911); and to the type of rock spalling that develops rounded residual boulders (Grabau, 1920, pp. 393-394; Blackwelder, 1925). Other names such as "concentric exfoliation," "concentric weathering," and "spheroidal weathering," have also been applied to the process by various authors. The present writers believe that the term spheroidal weathering is the most appropriate for this type of weathering, since it is expository and describes the type of boulder that is formed by the process. This name, therefore, will be used in the present paper.

During the last 90 years several hypotheses have been advanced to explain the spalling of rocks to produce rounded forms. It appears certain now, however, largely as a result of the careful studies of Eliot Blackwelder (1925, 1926), that spheroidal weathering results from the swelling and spalling of the exterior surface of blocks of rock due either to: (1) the hydration, carbonation, and oxidation of constituent minerals in the outer surface, or (2) the sudden intense heating of the outer surface by forest fires and brush fires. The senior author of the present paper has observed, however, in discussing the matter with others, that in spite of Blackwelder's work some geologists and many students are still confused by spheroidal weathering. Many fail to appreciate the relative importance of the two processes involved or are unable to discern which process operated in a particular case. The confusion is increased by the use of the name "exfoliation" for this type of weathering, a term that has also been applied to other somewhat similar processes. The purpose of the present paper is to clarify the subject and dissipate some of this confusion.

The first part of this paper is an historical review of the several hypotheses that have been advanced to explain the phenomenon of spheroidal weathering. The development of ideas is traced and the current views on the process are summarized. The second part of the paper is a description of the actual mineral alterations that have taken place in the shells surrounding several spheroidal boulders. It is believed that this information is a worthwhile contribution inasmuch as heretofore the mineral alterations accompanying spheroidal weathering have been described rather generally and vaguely, and, so far as the authors are aware, no detailed microscopic description of such changes has ever been published.

DEVELOPMENT OF HYPOTHESES ON THE ORIGIN OF SPHEROIDAL WEATHERING

Many years ago Page (1861, p. 75) in connection with his studies on the structures of igneous rocks, noted that some greenstones present a spherical or globular structure, the weathered cliffs of such a rock appearing like a huge accumulation of bombs and balls varying from a few inches to several feet in diameter. "Such a structure," he said, "from its apparent aggregation round a common centre, is also termed concretionary, and generally exfoliates, on exposure to weather, film after film, like the coats of an onion." Page evidently thought that the spheroids produced by weathering were fundamentally concretionary in nature and were merely brought out by weathering.

Bonney (1876) related that Professor J. Thompson regarded spheroidal "structure" as "the result of a process of exfoliation due to the action of the weather on a tolerably regular-shaped, homogeneous mass" (p. 149) along joint planes. Bonney disagreed with Thompson, however. He was willing to concede that cuboidal blocks of rock obviously tend to weather into rough spheroids; also that further action of the weather might occasionally produce concentric exfoliation in such spheroids; however, he did not believe that any theory of decomposition was adequate to explain the facts. Bonney believed that spheroids could form without the presence of joint planes. He supported his argument by citing the columnar basalt near Le Puy, Central France, in which "spheroids may be seen, . . . enclosed three or four at a time in a columnar shell without any dividing cross joints, so that they are just like Dutch cheeses packed in hexagonal cases (the interstices being filled up). The lid of the box has more or less fallen away and exposed the contained spheroids." (p. 150). He pointed out that since these spheroids are all solid, weathering has not occurred because there are no joints along which it can proceed. Bonney was convinced that spheroidal "structure" is due chiefly to contraction of the mass while cooling. During the discussion which followed the presentation of Bonney's paper before the Geological Society of London, a Mr. Koch told of an experiment in which he had placed fragments of ironstones and quartzite upon the sole of a furnace. These fragments cracked off in shells, leaving spheres. Thus Bonney's paper considers three hypotheses for the origin of spheroidal weathering: (1) Professor Thompson's theory of chemical decomposition caused by weathering along joint planes; (2) Bonney's theory that spheroids are the result of contraction of the rock mass while cooling; and (3) Koch's discovery that spheroids may be produced by the heating of rocks.

Geike (1882, p. 335) interpreted the spheroidal boulders of igneous rock as a result of weathering. He observed that: "In many prismatic massive rocks (basalt, diorite, etc) segments of the prisms weather into spheroids, in which successive weathered rings form crusts like the concentric coats of an onion. When one of these rocks has been intruded as a dike, it sometimes decomposes to a considerable depth into a mass of brown ferruginous balls in a surrounding matrix—the whole having a resemblance to a conglomerate made of rolled and transported fragments."

Along the Potomac River, outside of Georgetown, Washington, D. C., Spencer (1885) noticed perfectly rounded boulders of gneiss in a large mass of decayed rock of similar composition. The hillside had been cut away in the construction of a road beside the river, and exposed decayed crystalline rock to a depth of nearly 50 feet Much of the gneiss was disintegrated but contained unaltered masses which had resisted atmospheric decay Some of the gneiss upon weathering exhibited a schistose structure, yet much was remarkably compact, but traversed by numerous joint planes extending in all directions. As the weathering proceeded from the joint planes it left solid masses which were spheroidal and often showed a banded structure. Surrounding them were concentric zones which he decided marked the progress of decay and were in no way related to concretionary structure.

Thus far, geologists had noted spheroidally shaped boulders, enclosed in a series of concentric shells which they believed were caused by weathering, but they had not explained the mechanics of the process.

The importance of oxidation and decay in the production of spheroidal weathering was first clearly outlined by Dana (1895, p. 127). According to him: "This oxidation process, and other methods of decay go on with greatest rapidity in the fissures of rocks below a surface soil, because the descending surface waters keep them almost continuously wet; and it is under such circumstances that a rock which is much fissured or jointed becomes reduced to a pile of great boulders with a rusty earth between. The decay of oxidation at first produces a thin discoloring of adjoining surfaces, and this continues, eating off the angles, which are attacked from three directions, until a bluff of solid rock becomes apparently a pile of great boulders. With the progress of alteration, the discolored portion becomes banded with yellow and brown; and as it deepens, the outer part of the spheroid sometimes separates in concentric shells, precisely corresponding with the concentric structure of a concretion. But these concentric shells are due to the decay that is in progress; and apparently to alterations in the work of decay dependent on climate . . . " Dana thus recognized that this spheroidal weathering is due to a definite type of chemical action related to climatic conditions.

The "niggerheads" of the gabbro area around Baltimore, Maryland, were cited by Merrill (1897, pp. 241-248), as typical examples of spheroidal weathering. These massive rocks, he thought, had been originally traversed by one or more sets of joints, along which moisture and "the accompanying agents of disintegration" had made their way and gradually rounded off the corners until only an oval mass remained, surrounded by concentric layers. Merrill believed that the above process holds true also for huge granite bosses, or domes, such as those in Yosemite Valley, California, and Stone Mountain, Georgia. These domes, he felt, were merely on a much grander scale. He did not elaborate upon the mechanics of the process.

Some years later J. F. Kemp (1909) received from a Mr. J. R. Villars a photograph accompanied by a letter discussing the curious weathering of a diabase dike near Butte, Montana.

This dike, which was exposed to a depth of 20 or 30 feet, showed rounded masses of rock varying in diameter from a few inches to two feet. These masses were separated into concentric layers, a division which was marked in the weathered rock but was wanting or scarcely observed in the fresh rock. Kemp forwarded the photograph and letter to the Mining and Scientific Press stating that: "The dike is a fine illustration of spheroidal weathering . . . the breaking up being in large degree referable to strains produced in cooling. Weathering has brought out the shelly or onion structure, but one cannot well resist the conviction that some internal structure has facilitated the assumption of these peculiar shapes." (p. 443). Although Kemp's conclusions are not at all clear, it appears that he attributed spheroidal weathering of the diabase to a combination of three factors: (1) internal structure, (2) strains caused by cooling, and (3) weathering.

Hobbs (1912, pp. 150-152) was one of the first to recognize two distinct processes in the development of rounded boulders resulting from spheroidal weathering: (1) mechanical disintegration, and (2) chemical decomposition. Hobbs believed mechanical disintegration is due to daily temperature changes. Relative to chemical decomposition he made an effort to analyze its progress. He stressed the importance of jointing in facilitating the progress of chemical decay which he thought was brought about by the process of hydration and carbonation. He emphasized that the newly formed minerals, resulting from hydration and carbonation, are notably lighter and hence more bulky than those minerals from which they have formed. "Strains are thus set up," he argued, "which tend to separate the bulkier new material from the core of unaltered rock. Eventually, the squared block is by this process transformed into a spheroidal core of still unaltered rock wrapped in layers of decomposed material . . . " (p. 151).

Cleland (1916, pp. 31-33, 39-40, 388), like Hobbs, believed that rounded boulder-like forms can be developed by two distinct processes of weathering. One of these he termed "spheroidal weathering" which he considered to be brought about by chemical decomposition occasioned by hydration. The other he called "exfoliation" and explained it as mechanical disintegration brought about by daily changes in temperature.

Grabau (1920, pp. 393-394) believed that rounded boulders are produced in two ways: (1) by "concentric exfoliation" whereby the rock peels off in concentric shells, and (2) by granular disintegration whereby the rock crumbles into grains but leaves a rounded residual boulder. "Concentric exfoliation" he attributed to daily changes of temperature. These changes, he maintained, are especially great in desert regions and have been estimated on the surface of some rocks to be as great as 80°C. The rock, being a poor conductor of heat, is affected chiefly on the surface, although the heat passes inward very gradually. On cooling by radiation in the clear atmosphere, the temperature of the outer layers sinks rapidly and easily passes below that of the inner part of the rock. As a result, the surface portion, for some distance inward, is subjected to a series of stresses which results in the flaking or peeling off of the outer layers. Since the angles of the rock are most exposed, being subjected to heating and cooling from all sides, they peel off first and the result is the production of curved and rounded outlines. Thus, according to Grabau, "concentric exfoliation" or the successive peeling off of layers results, a phenomenon very marked in most dark and fine-grained igneous rocks in desert regions. Grabau illustrates this process by a photograph showing what he calls "concentric exfoliation" in a dike of basic igneous rock taking place at a depth of at least 25 feet below the surface. Most people today would agree, however, that temperature changes are insignificant at such depth. Furthermore, the surface of the ground shown in this photograph is covered by rather dense vegetation, indicating that the climate is reasonably moist and not arid as intimated.

The other process that Grabau believed effective in the development of rounded boulders is granular disintegration, which he attributed also chiefly to daily temperature changes. This process may operate in three different ways: (1) Since different minerals, such as plagioclase and pyroxene, for example, have different coefficients of expansion and contraction, internal stresses are set up in the rock which tend to separate these minerals, one from the other, until a loose sand results. (2) Individual minerals themselves are affected by temperature changes due to the fact that they expand and contract differently in different crystal directions. The stresses thus set up tend to open up minute cracks along cleavage planes into which air and moisture can penetrate and effect decomposition of the mineral. (3) "In moist climates, dark igneous rocks are often reduced to a mass of residual boulders by combined disintegration and decomposition." (p. 397).

The first systematic study of spheroidal weathering was made by Blackwelder (1925) who used the term "exfoliation" for the process. He pointed out that exfoliation, or the development of spheroidal forms, takes place most commonly in the granitoid igneous rocks, and occurs under various conditions. Blackwelder read some 30 standard reference works and special papers on weathering and decided that the majority of writers attributed exfoliation to the daily range of temperature in deserts and on mountain tops where the range is exceptionally great. As a result of intensive field work, however, he concluded that exfoliation is not a simple process, but a combination of processes of diverse origins.

In typical exfoliation, according to Blackwelder, "a series of cracks is formed roughly concentric to a single joint-block . . . The force producing the cracks has been one of tension, with the effective component operating in a radial direction . . . Such radial forces originate either by (a) the internal shrinkage of the block (b) by its external expansion" (p. 794).

In order to test the hypotheses of some writers that temperature changes cause exfoliation, Blackwelder made laboratory experiments in which he subjected sound igneous rocks to sudden changes of temperature, ranging from 15° to 210°C., by plunging the cold rock into boiling oil. In no case did these cause any spalling or cracking or even visible weakening of the rock. Inasmuch as this laboratory range in temperature is greater than any possible temperature range in the desert, the apparent lack of effect is significant.

From field observation Blackwelder noted that rocks in very dry deserts are rarely exfoliated. He examined hundreds of boulders strewn over the great alluvial fans in the Panamint and Salton deserts of California. He climbed many mountain peaks, 1000 feet to 3000 feet above timber-line, but found no evidence of exfoliation. In the humid climates, however, where the diurnal range of temperature is less than in the desert, he found that exfoliation appears abundantly in outcrops of granite, gabbro, and diabase. Furthermore, in these moister regions he found boulders of exfoliation forming at a depth of

20 feet below the surface. He considered these observations as cogent evidence that exfoliation is due to chemical decay.

A year later Blackwelder (1926) pointed out that in the semi-arid forested mountains of western United States fire seemed to rank first in causing the disruption of boulders and rock outcrops. This statement was supported by a series of experiments in which Blackwelder subjected different kinds of igneous rocks to repeated sudden heatings and coolings with temperatures ranging from 200° to 800°C. There was no visible effect upon the rock unless it approached the extreme temperature range that the rock was capable of withstanding. From these experiments Blackwelder concluded that insolation, or diurnal changes of temperature, are entirely inadequate to cause rock breakage. However, since the temperatures of fires are likely to reach 1200°C. and even above, he believed that forest fires can cause rocks to spall. He stated: "The field evidence is abundant, but has received little notice. On the broad plateau of the Medicine Bow Mountains, as well as in the Sierra Madre and other ranges in southern Wyoming, the writer has found nearly every boulder and outcrop in the forested zone thus cracked and rounded and the adjacent ground strewn with spalls. The effect of all other weathering processes were insignificant in comparison with those of the repeated conflagrations. Similar conditions were noted along the semi-arid eastern flank of the Sierra Nevada in California. In such localities it is probable that fire spalling demolishes exposed rocks much more rapidly than the more common but less violent processes such as chemical disintegration." (p. 138-139).

Like Blackwelder, Grout (1932, pp. 305-307) believed that the change in rock volume caused by daily and seasonal temperature variations is small in comparison with that resulting from forest fires, though the alterations caused by the former are more numerous and may have an effect over a period of years. He believed that: "the spalling produced by a single forest fire is probably comparable to the effects of sun and frost for a thousand years, though the great effect of fire may be partly a result of previous loosening of the grains by frost and other agencies."

The experiment by Griggs (1936) was designed to test the effectiveness of fatigue in rock exfoliation by insolation. Although this experiment dealt with the process of exfoliation in the general sense, the results are significant with respect to spheroidal weathering. Griggs subjected a specimen of granite to artificial temperature changes in a 15-minute cycle varying from 32°C. to 142°C. This experiment was continued for the equivalent of 244 years of diurnal temperature change. Photomicrographs of the surface of the rock before and after the experiment showed that there had been no cracking or spalling as a result of heating and cooling. Griggs concluded that "the effect of temperature changes over a thousand years is not sufficient to cause any exfoliation or disintegration of granite" (p. 796).

The importance of fire was further emphasized by Emery (1944) who investigated the relations of brush fires to rock exfoliation. He studied several square miles of brush and grass land in San Diego, California, where he found that exfoliated rock outcrops are numerous. He observed that: "Spheroidal shapes are characteristic of quartz diorite outcrops in the region and further development of these shapes resulted from the exfoliation by the fire, since the thickest spalls were generally formed at the sharpest corners. Loosely attached to a few of the rocks were older but very similar spalls, suggesting that the spheroidal shapes of the rocks were partly due to previous brush fires . . . Since the heat even of a brush fire can produce extensive exfoliation, and, since large areas of the west have a brush cover, it seems not unlikely that a great deal of exfoliation has resulted from brush fires, with the consequent tendency toward development of spheroidal rock shapes in widespread areas" (p. 508).

Very recently Larsen (1948, pp. 114-119) described and illustrated some granitic rocks in Southern California that have weathered into rounded forms which he called "boulders of disintegration." These boulders are commonly buried 10 to 20 feet below the surface where they are imbedded in a gruss of disintegrated rock. The boulders themselves are fresh and some are surrounded by shells of fresh or partly decomposed rock. Many boulders now project above the surface or lie upon the surface. Larsen believes that at one time these were imbedded in gruss and later the intervening gruss was washed away leaving the boulders on the surface. He made chemical analyses of the fresh boulders and of the gruss surrounding them and from

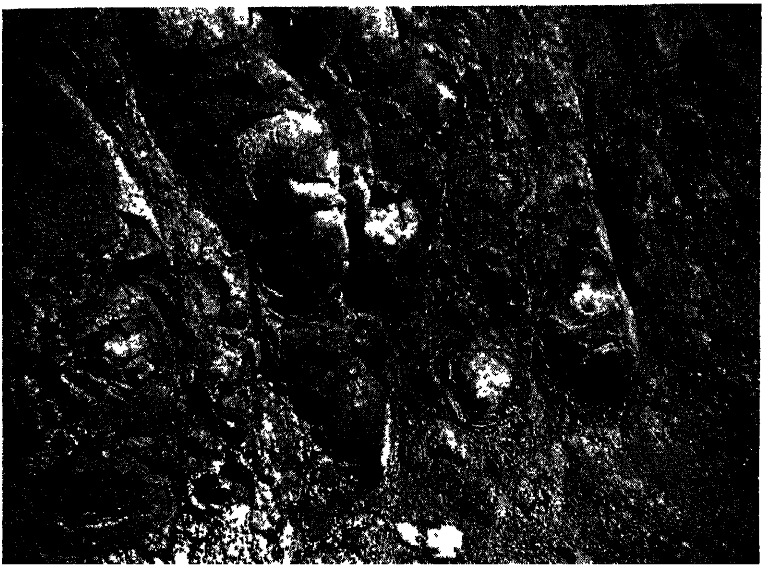


Figure 1 Spheroidal weathering in porphyritic olivine basalt at the north end of the Koolau Range, Oahu This exposure is in a steep cut and about 30 feet below the ground surface Note that the spheroids are arranged more or less regularly in vertical and horizontal rows determined by joint planes Some vertical joints are still visible The spheroid in the extreme lower left corner is about 18 inches in diameter

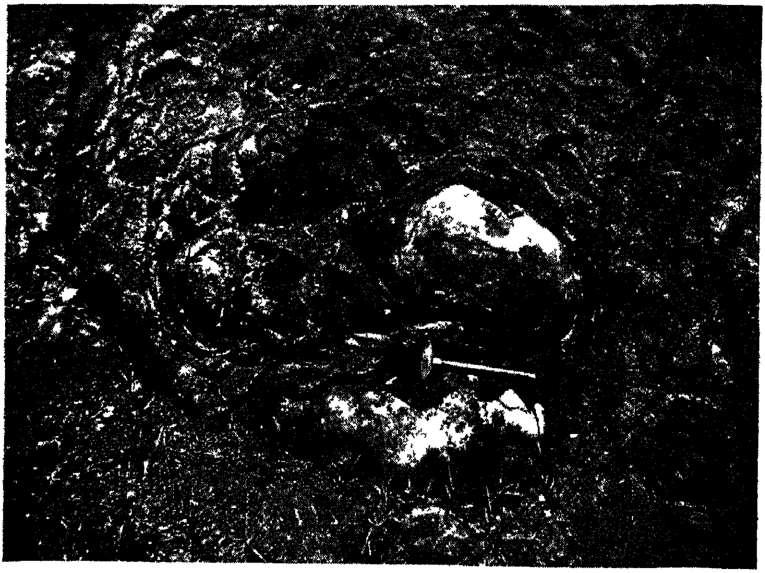


Figure 2 Detailed view of spheroidal weathering of porphyritic olivine basalt at the north end of the Koolau Range, Oahu Note how the firm, unaltered rock in the interior of the spheroids grades radially outward through weathered shells into completely disintegrated rock

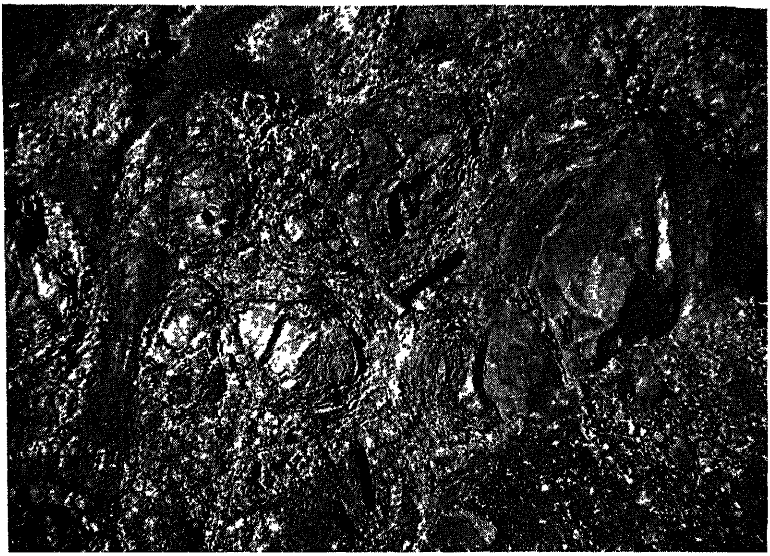


Figure 3 Spheroidal weathering in vesicular basalt on the Schofield Plateau near the west gate of Wheeler Field, Oahu This outcrop is in a road cut about 12 feet below the ground surface Note how weathering has taken place along joints which are less regular than in Figures 1 and 2

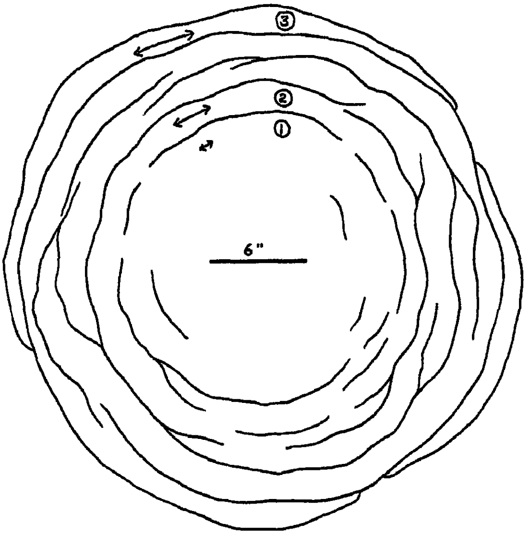


Figure 4 Hypothetical cross section through the boulder of porphyritic olivine basalt described in the text, showing the approximate location of the three specimens chosen for microscopic study Specimen 1 is essentially unaltered except for slight serpentinization of the olivine phenocrysts Specimen 2 is more weathered, with labradorite crystals altering to montmorillonite, and rims of iddingsite around olivine phenocrysts changing to limonite Specimen 3 is badly weathered, labradorite crystals have been completely converted to montmorillonite and augite crystals and iddingsite rims are clouded with limonite The arrows indicate the direction and relative intensity of the expansive force in each layer resulting from alteration of the rock These expansive forces have caused the rock to scale

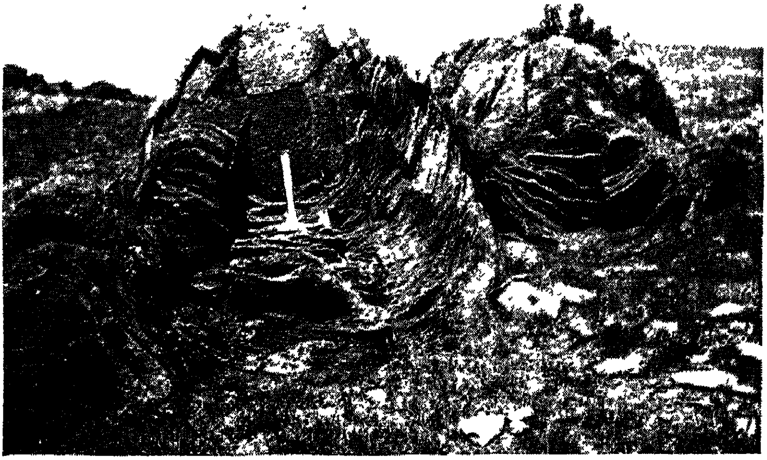


Figure 5 Spheroidally weathered boulders of biotite granite near U S Highway 287, about 5 miles south of Virginia Dale, Colorado Although this is a semi-arid region, temperature variations are not extreme and the shelling of the exterior is due to decomposition and not to temperature changes

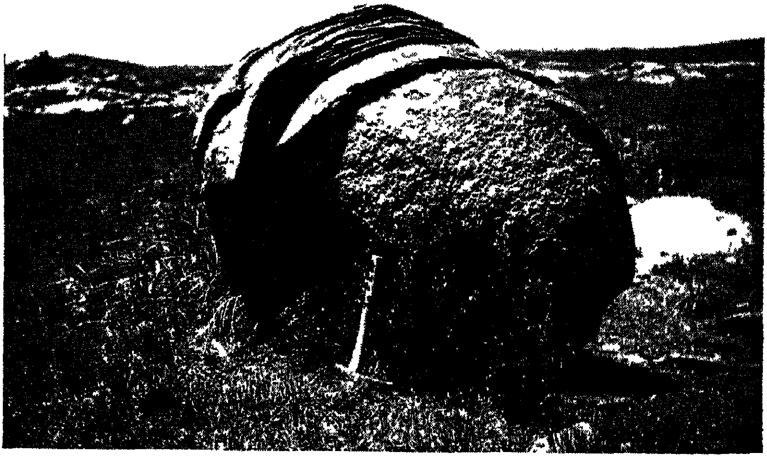


Figure 6 An especially fine specimen of spheroidally weathered biotite granite along U S Highway 287, about 5 miles south of Virginia Dale, Colorado This boulder is a residual that remains after the surrounding rock has weathered away Microscopic examination reveals that the minerals in the outer shells are more altered than those in the inner ones and that spalling is due to chemical decay

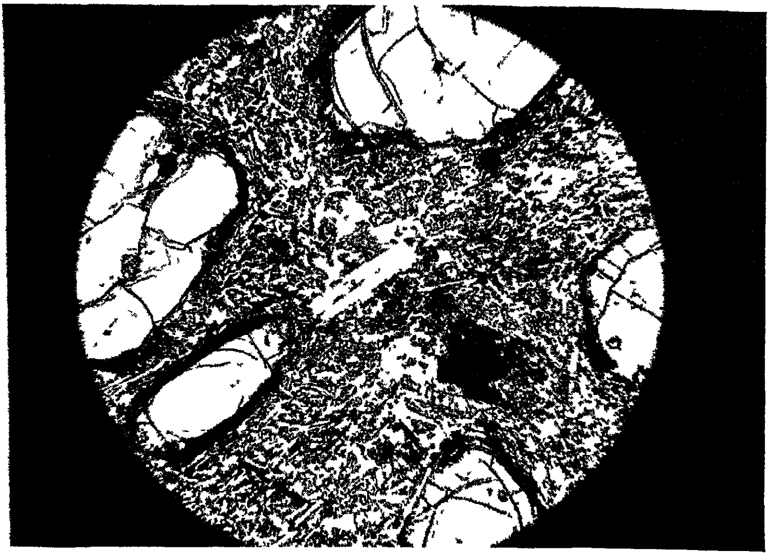


Figure 7 Photomicrograph of slightly weathered porphyritic olivine basalt (specimen 2) from the innermost shell of a spheroidally weathered boulder This specimen shows slightly more alteration than the fresh basalt (specimen 1) of the interior core The iddingsite rims about the large phenocrysts of olivine have become nearly opaque by alteration to limonite In the groundmass the crystals of augite (high relief) and magnetite (black) remain fresh, but the centers of many labradorite crystals (white and lath-shaped) have altered to montmorillonite Photomicrograph taken in plane polarized light Diameter of field about 3 mm

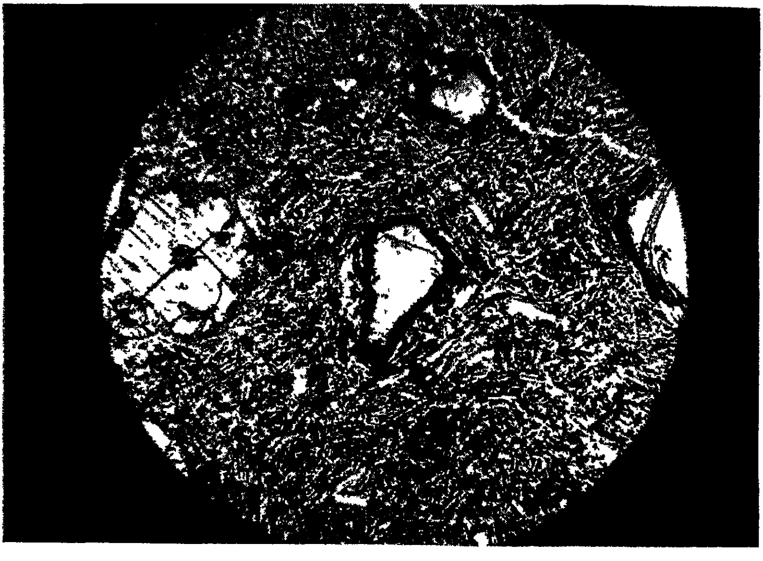


Figure 8 Photomicrograph of badly weathered porphyritic olivine basalt (specimen 3) from the outermost shell of the same spheroidally weathered boulder mentioned in Figure 7 The iddingsite borders on the phenocrysts of olivine are completely oxidized to limonite The major alteration has been in the groundmass where the augite crystals are now permeated with tiny specks of limonite, making them nearly opaque The labradorite crystals, although retaining their outlines, have been converted to a lamellar aggregate of montmorillonite Photomicrograph taken in plane polarized light Diameter of field about 3 mm

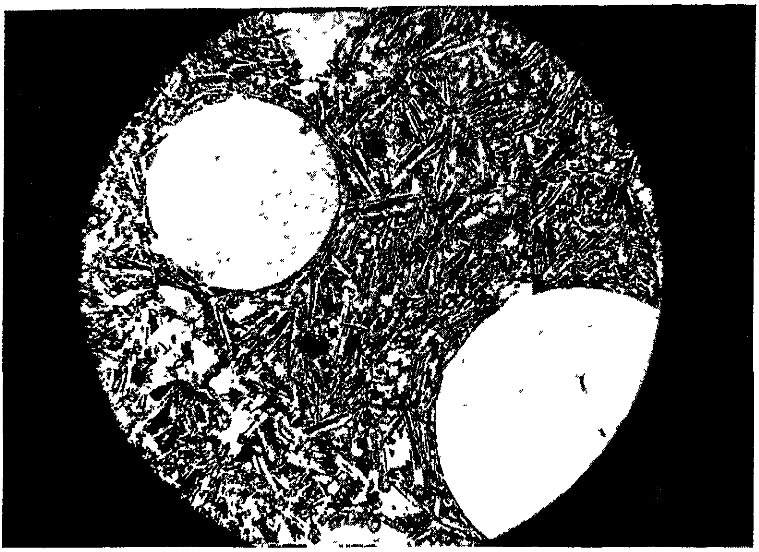


Figure 9 Photomicrograph of relatively fresh vesicular basalt (specimen 4) from the periphery of the interior core of a spheroidally weathered boulder The rock consists of well-rounded vesicles and a groundmass of labradorite (white laths, some with dark interiors), augite (somewhat clouded, irregular and elongate grains with high relief), magnetite (black grains), and interstitial glass (small colorless areas) The interiors of some labradorite laths are slightly saussuritized and some augite crystals are clouded by limonite. Photomicrograph taken in plane polarized light Diameter of field about 8 mm

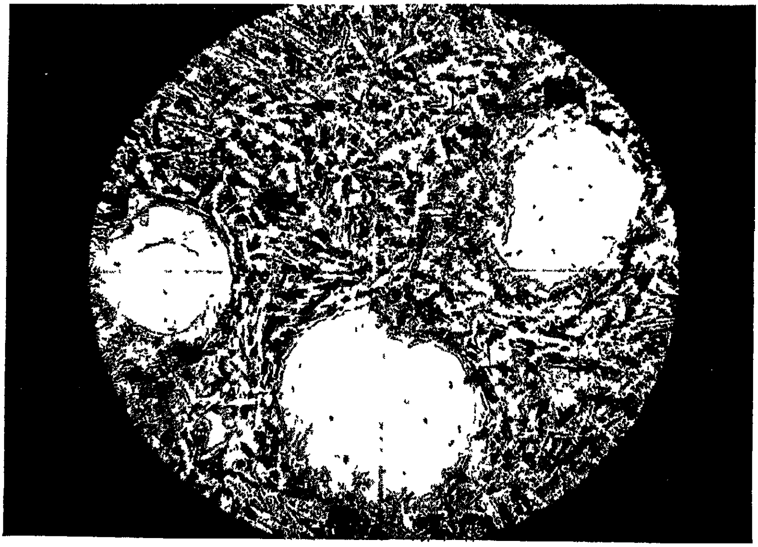


Figure 10 Photomicrograph of weathered vesicular basalt (specimen 5) from the outer shell of the same spheroidally weathered boulder mentioned in Figure 9 Here the augite crystals are badly decomposed and heavily permeated with specks of limonite The labradorite crystals, although retaining their outlines, are almost completely converted to montmorillonite Magnetite has partly altered to hematite. The walls of the vesicles have been deformed and pushed inward, apparently by the swelling of the rock through alteration Photomicrograph taken in plane polarized light Diameter of field about 8 mm.

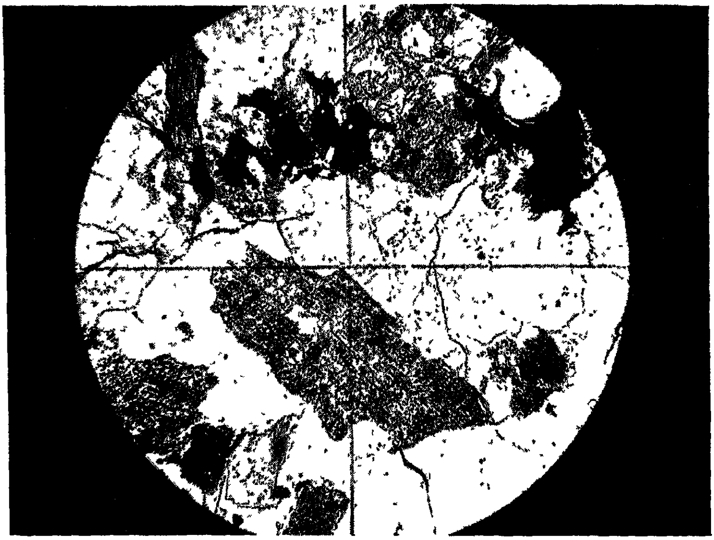
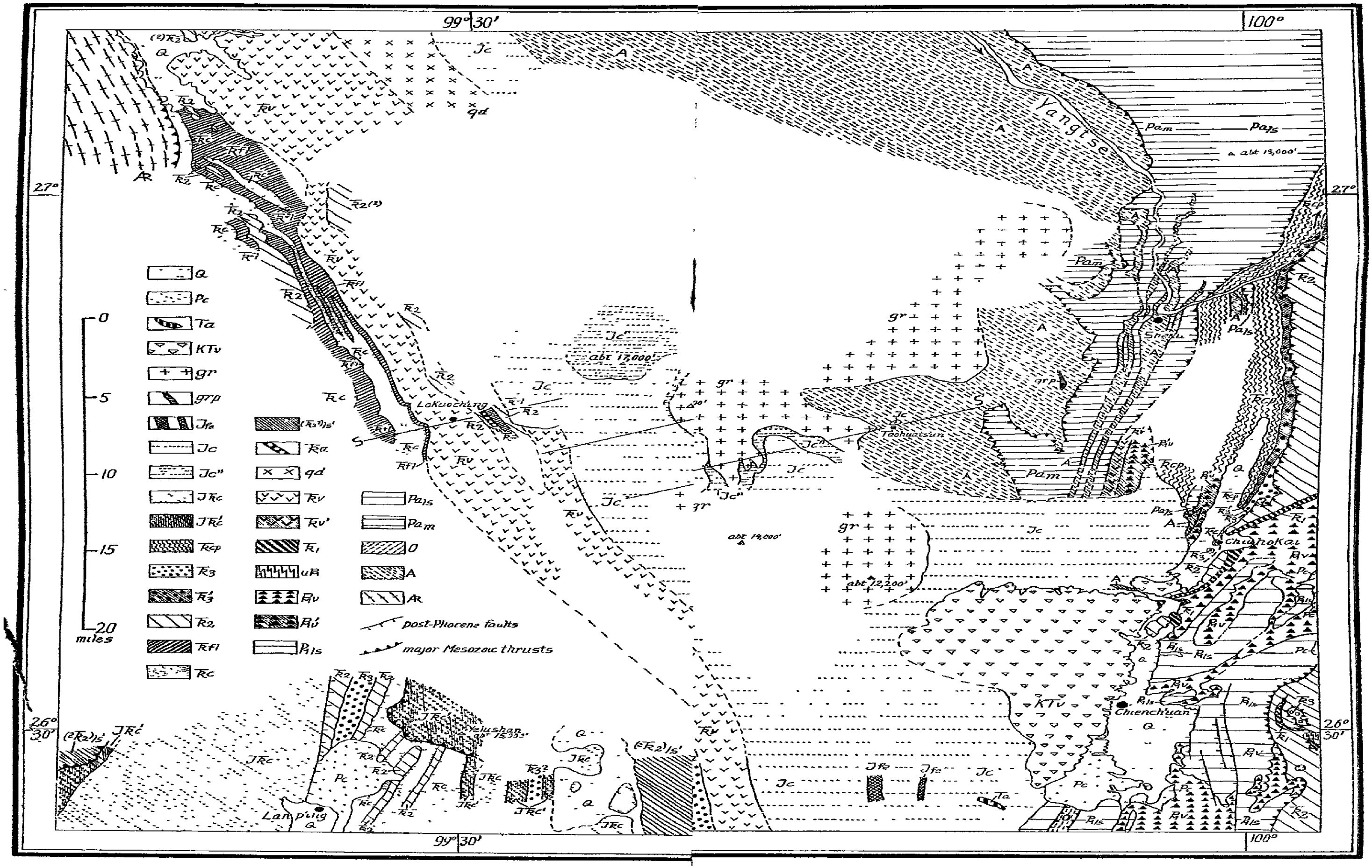


Figure 11 Photomicrograph of slightly weathered biotite granite (specimen 7) from an outer shell of a spheroidally weathered boulder The white unaltered minerals are chiefly quartz and some microcline The gray minerals, including the large crystal near the center, are oligoclase, which is deeply clouded with kaolinite and coated with large flakes of sericite (clear areas) The black grains are magnetite, and the dark gray mineral at the lower right edge of the photograph is biotite altering to chlorite In the oligoclase of this section the flakes of sericite are larger and more abundant, and the kaolinization is more intense than in the oligoclase of the relatively fresh biotite granite (specimen 6) (photomicrograph not shown) Photomicrograph taken in plane polarized light Diameter of field about 3 mm


99° 30'
100°
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26° 30'
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Lokuochiang
Taohuats'un
Shehu
Chunhokai
Chienchuan
Lanping
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post-Pliocene faults
major Mesozoic thrusts

the results concluded that the "boulders of disintegration" were formed by the chemical weathering of the rock and its accompanying disintegration.

SUMMARY OF CURRENT VIEWS ON SPHEROIDAL WEATHERING

From the foregoing it is clear that most writers now consider spheroidal weathering to be caused by either of two processes: (1) the swelling and spalling of the exterior surface of angular blocks of rock due to the hydration, carbonation, and oxidation of constituent minerals in the outer surface, or (2) the swelling and spalling of the exterior surface of angular blocks which have been subjected to sudden, intense heating of their exterior by forest fires and brush fires.

The first process, which may take place in both humid and semi-arid climates, is the more common and operates not only at the surface of the ground but in some cases to a depth of 30 or 40 feet. The other process is undoubtedly local and confined to the surface of the ground since appreciable temperature changes by fires or by other means are not effective at depth. Accordingly, where spheroidal boulders are found forming at depth, it can be safely concluded that the process is one of hydration, carbonation, and oxidation of the minerals in the outer layers and not one of temperature changes. Spheroidal boulders lying on the surface may be the result of either of the two processes. Although it is assumed by many that such boulders located in semi-arid regions are the result of temperature changes, probably most have been caused by hydration, carbonation, and oxidation of minerals in the outer layers. Boulders of this type are well illustrated in Figures 5 and 6 and will be described later in this paper.

MICROSCOPIC STUDY OF SPHEROIDALLY WEATHERED BOULDERS

GENERAL STATEMENT

In the previous section it was concluded that the principal cause of spheroidal weathering is the expansion of the outer surface of a rock due to the hydration, carbonation, and oxidation of its mineral constituents. Although this explanation is now used by many writers, it is stated rather generally and vaguely without specific reference to the exact nature of the mineral changes. As far as the present writers are aware, no microscopic study has ever been made of spheroidal boulders

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Three rocks exhibiting particularly good spheroidal weathering were selected for study (figs. 1 to 6). One of these is a porphyritic olivine basalt from the northern end of the Koolau Range, Oahu, T. H.; the second, a vesicular basalt from the Schofield Plateau, near Wheeler Field, Oahu; and the third, a biotite granite from northern Colorado. In each case an especially good spheroid was chosen and a series of specimens was collected along a radius to represent the fresh rock of the boulder itself and the more weathered rock of the shells surrounding it. Thin sections were prepared from these specimens and mineral alterations were studied (figs. 7 to 11). Modes were determined with a Wentworth integrating stage and grain sizes measured with a micrometer ocular. Microscopic study of these specimens has enabled the writers to draw certain significant conclusions about the relationship of mineral alterations to spheroidal weathering.

PORPHYRITIC OLIVINE BASALT

General nature. This rock, belonging to the Koolau volcanic series, outcrops at the northern end of the Koolau Range, Oahu, about 3 miles southwest of the village of Kahuku. At this point about 1,000 feet above sea level, the lavas have been deeply dissected by streams, and one section is well exposed in a steep cut nearly 30 feet high where an old road follows a canyon wall. The basalt is so thoroughly weathered into spheroidal masses that the outcrop simulates a sedimentary deposit of rounded boulders (figs. 1 and 2). Spheroidal weathering extends from the surface to a depth of at least 30 feet.

The spheroids are lined up in more or less regular rows, arranged approximately vertically and horizontally, and individual spheres are spaced on centers ranging from 8 inches to 3 feet. At the contact between any two spheroids the line of demarcation is generally planar and suggests that a joint plane served as a starter. In many places the original joints are still visible. The spheroids themselves range from 6 inches to 3 feet in diameter, and are larger and less weathered near the base of the cut. They are bounded by concentric shells which are concavo-convex with a maximum thickness near their centers ranging from half an inch to 3 inches, and which taper toward their edges.

The interior of the spheres is composed of fresh, firm rock consisting of a blue-green groundmass set with rounded, yellow-green phenocrysts of olivine. As one progresses from the fresh rock of the interior toward the outer shells, the groundmass becomes brownish yellow and eventually the olivine turns brown, loses its luster, and falls out of the specimen. The final weathering product surrounding the outer shells is a buff, brown, or red, friable rock which crumbles to a powder. The complete zone of alteration, i.e., from fresh rock to completely weathered material, takes place within a zone from 3 to 6 inches wide.

One particular spheroid, 2½ feet in diameter, was selected for microscopic study. The solid interior is 9 inches in radius and its rim of concentric shells is 6 inches thick (fig. 4). Three specimens were collected along the radius of the spheroid as follows: (1) fresh basalt (specimen 1) from the outer edge of the solid interior, 8 to 9 inches from the center; (2) slightly weathered basalt (specimen 2) from the innermost shell, 9 to 10 inches from the center; and (3) badly weathered basalt (specimen 3) from the outermost shell, 14 to 15 inches from the center.

Fresh basalt (specimen 1). This is a porphyritic rock consisting of a dense, light-gray groundmass set with abundant sub-rounded phenocrysts of yellow-green olivine. In thin section it was found to consist of sub-rounded, embayed phenocrysts of olivine and much smaller microphenocrysts of hypersthene, set in a dense groundmass of labradorite, augite, magnetite, and volcanic glass. Many of the olivine phenocrysts contain alteration rims of yellow-brown iddingsite averaging 0.4 mm. thick. The microphenocrysts of hypersthene have an average diameter of 0.22 mm. and all appear to have crystallized later than the olivine. The labradorite ($Ab_{40}An_{60}$) forms slender laths 0.1 mm. long which show random orientation except adjacent to olivine phenocrysts where they trend parallel to the borders. Small rounded grains of augite and magnetite are scattered among the labradorite, and colorless, slightly birefringent volcanic glass fills the interstices between grains. The mode of this rock, in volume per cent, determined with an integrating stage, is shown in Table I together with the grain sizes of the constituent minerals.

TABLE I

Modes, in volume per cent, and grain sizes of porphyritic olivine basalt.

Minerals	*Modes*		*Grain Sizes*	
	Specimen 1	*Specimen 2*	*Range*	*Average*
Olivine	47.1%	46.9%	0 4–3.5 mm.	1.8 mm.
Hypersthene	8.7	9.5	0.1–0.3 "	0 2 "
Augite	15.8	16.9	(uniform)	0.02 "
Labradorite (ca. An_{60})	21.3	19.8	"	0.1 "
Magnetite	1.7	2.0	0.02–0.2 mm.	0.04 "
Glass	5.2	4.9	——	——
Miscellaneous	0.2	——	——	——
	100.0%	100.0%		

Except for slight serpentinization of some olivine phenocrysts along fractures, the rock is essentially unaltered. The labradorite, augite, and magnetite of the groundmass are quite fresh. The development of iddingsite rims is believed to be a late magmatic phenomenon.

Slightly weathered basalt (specimen 2). In this specimen the groundmass has turned light gray or buff and some of the olivine phenocrysts have lost their brilliant luster. The rock is less firm than specimen 1 and crumbles readily under a blow of the hammer. In thin section (fig. 7) this specimen is texturally similar to specimen 1 and its minerals are qualitatively similar (table I). The apparent small quantitative mineral differences between the two rocks are due partly to limitations in accuracy of the Rosiwal method and partly to actual original differences in the two specimens. The volcanic glass in specimen 2 is olive-green instead of colorless, possibly due to weathering. Tiny acicular microlites penetrate the glass.

This specimen shows slightly more alteration than specimen 1. All iddingsite rims about the olivine have been partly altered to limonite and have become nearly opaque. The olivine itself, however, remains relatively fresh as do the augite and magnetite of the groundmass. The centers of many labradorite laths have been altered to a fine-grained aggregate with low birefringence which is believed to be montmorillonite, but the borders of the laths remain fresh.

Badly weathered basalt (specimen 3). In this specimen the groundmass has become yellow-brown. Although most olivine phenocrysts retain a high luster internally, their surfaces have turned dull and brown. These phenocrysts can be easily removed with the fingers and the rock can be crumbled in the hand. In thin section (fig. 8) the texture is found to be similar to that in the previous specimens and it is apparent that the original mode was also the same (table I).

This rock is highly altered. Iddingsite borders on the olivine phenocrysts are completely oxidized to limonite, but the olivine itself is still relatively fresh although slightly more serpentinized than in specimen 2. The major alteration has taken place in the groundmass. Augite crystals have been weathered and are now permeated with tiny grains of limonite which effectively obscure the crystals and prevent more thorough optical study. Labradorite crystals have been intensely altered throughout to a lamellar aggregate of montmorillonite, and only locally may the original laths be recognized. Most magnetite granules appear completely unaltered.

VESICULAR BASALT

General nature. This rock, a fine-grained, bluish gray, vesicular basalt of the Koolau volcanic series, is exposed in a deep road cut on the Schofield Plateau near the west gate of Wheeler Field, Oahu. The cut reveals spheroidal weathering of the basalt to a depth of 12 feet below the surface (fig. 3).

In general the spheroids are spaced on centers from 1 to 3 feet apart, but are not as well aligned as those in the porphyritic olivine basalt in the Koolau Range. The spacing of spheres and the arrangement of shells of adjacent spheres show clearly that the weathering began along joint planes. The shells are thinner than those in the basalt from the northern end of the Koolau Range. Generally they range in thickness from an eighth of an inch to 2 inches but the average is between a quarter and three-quarters of an inch. Furthermore, the thickness of an individual shell is more uniform throughout than in the porphyritic olivine basalt and the shell extends farther around the interior boulder; thus the shell is less lenticular and the interior boulder is more completely surrounded by it.

The interior residual boulders are fresh and unjointed. The first apparent evidence of weathering is seen in the innermost shells that have maintained their texture but have turned grayish. In the outer shells the rock turns brown to yellow-brown and starts to fall apart. Surrounding the outer shells is a zone of brown or red-brown, completely decayed rock. The width of the weathered zone, i.e., from the fresh rock to the completely decayed rock, ranges from 6 to 12 inches.

Two specimens of this basalt were selected for microscopic study from one spheroidal mass, 2 feet in diameter: (1) relatively fresh basalt (specimen 4) from the periphery of the interior spheroid, 5 to 6 inches from its center; and (2) weathered basalt (specimen 5) from an outer shell surrounding the solid interior, 11 to 12 inches from its center.

Relatively fresh basalt (specimen 4). This specimen is medium gray, aphanitic, and contains abundant spherical vesicles averaging 1 mm. in diameter. Microscopically (fig. 9) its texture is intersertal, the rock consisting of a mat of labradorite laths (An_{64}) and augite and magnetite granules with colorless to brownish basaltic glass filling interstices. The labradorite is randomly oriented except adjacent to vesicles where the expanding gas bubbles forced the adjacent laths into a concentric arrangement. The labradorite shows albite twinning and conspicuous zoning. Augite and magnetite granules are scattered among the plagioclase. Many of the augite grains are long and slender like the labradorite, with an average length of 0.1 mm., but some are equidimensional with an average diameter of 0.04 mm. Irregular grains of reddish-brown iddingsite, developed from olivine, are scattered throughout the rock, and in some of these, cores of olivine still remain. A few microphenocrysts of hypersthene were noted. The mode of this rock and the grain sizes of its mineral constituents are shown in table II.

Although relatively fresh, this rock shows two effects of weathering. In the first place, many labradorite cores are altered to a very fine-grained birefringent aggregate which appears to consist of sericite, kaolinite, and epidote, although the positive identification of these minerals is not certain. In the second place, some augite crystals have become brownish by oxidation to limonite.

TABLE II.

Modes, in volume per cent, and grain sizes of vesicular basalt.

Minerals	*Modes*		*Grain Sizes*	
	Specimen 4	*Specimen 5*	*Range*	*Average*
Vesicles	23.0%	22.3%	0.3–2.1 mm	1.1 mm.
Augite	22.6	24.1	0.02–0.2 "	0.1 "
Labradorite (An_{64})	33.5	47.9*	0.1–0.4 "	0.2 "
Glass	14.1		——	——
Magnetite	6.3	5.4	——	0.4 "
Olivine (altering to iddingsite)	0.5	0.3	0.04–0.3 mm.	0.1 "
			——	——
Hypersthene	tr.	——		
	100.0%	100.0%		

* The labradorite has been almost completely altered to montmorillonite. The ratio of montmorillonite to glass is difficult to determine.

Weathered basalt (specimen 5). The vesicular basalt from the outer shell of the weathered boulder has turned pinkish or yellowish and can be crumbled readily between the fingers. Under the microscope (fig. 10) it is texturally like the fresher material and its mode was probably originally about the same (table II). The rock has been conspicuously altered by weathering, however. Augite crystals are badly decomposed and have lost their birefringence, and many are now heavily permeated with specks of limonite. In some crystals serpentine appears to have developed but the clouded character makes this identification somewhat questionable. The labradorite crystals have been almost completely converted to lamellar aggregates of montmorillonite, and only locally do relics of the original mineral remain. In many places, however, the outlines of the laths are preserved. The volcanic glass can still be recognized but its ratio to montmorillonite is difficult to determine. Magnetite has partly altered to hematite.

The vesicles are not as smoothly circular as those in specimen 4. Practically all have been deformed and crystals of augite, altered labradorite, and magnetite have been pushed inward, forming irregular projections. It appears that the walls of the vesicles have been jammed inward by the swelling of the rock through alteration.

BIOTITE GRANITE

General nature. Many residual boulders of spheroidally weathered pre-Cambrian biotite granite lie upon the surface in northern Colorado (figs. 5 and 6). One particularly fine example (fig. 6), along U. S. Highway 287 about 5 miles south of Virginia Dale, Colorado, was chosen for study. This boulder, approximately 3 feet in diameter, has a solid interior surrounded by from 5 to 10 concentric shells ranging in thickness from 1 to 4 inches. For the most part the shells have uniform thickness throughout and appear to have completely surrounded the interior sphere at one time.

The interior boulder is composed of pink, medium-grained, non-foliated granite, without joints. The rock in the shells is deeper pink and somewhat more crumbly, but is otherwise megascopically similar to the interior. Two specimens were selected from microscopic study from this spheroidal body: (1) relatively fresh granite (specimen 6) from the innermost shell, about 10 inches from the center of the spheroid; and (2) slightly weathered granite (specimen 7) from an outer shell, 6 to 8 inches outward radially from specimen 6.

Relatively fresh granite (specimen 6). This specimen, from the innermost shell of the weathered boulder, is a pink, medium-grained granite, relatively fresh and strong. Microscopically it is allotriomorphic and generally even-grained, although some portions tend to be seriate. One large perthite phenocryst 6.3 mm. long was seen. Perthite, microcline, oligoclase, quartz, and biotite are the essential constituents and magnetite is an accessory. Sericite and kaolin have formed from the feldspars. The mode of this specimen in volume per cent, together with the grain sizes of the modal minerals, is shown in table III.

TABLE III

Modes, in volume per cent, and grain sizes of biotite granite.

Minerals	*Modes*		*Grain Sizes*	
	Specimen 6	*Specimen 7*	*Range*	*Average*
Perthite and microcline	36.5%	41.0%	0.07–2.1 mm.	1.0 mm.
Oligoclase (An_{10-14}) .	23.1	25.2	“ “ “	“ “
Quartz	31.1	24.6	“ “ “	“ “
Biotite	6.2	4.7	——	0.7 “
Sericite	1.6	3.3	(see note below*)	
Magnetite	1.5	1.2	——	0.09 “
Apatite	—	tr.	——	——
	100.0%	100.0%		

* In specimen 6 this mineral ranges from tiny specks to flakes 0.4 mm. across, with an average diameter of 0.1 mm. In specimen 7 it ranges from tiny specks to flakes 0.6 mm. across, with an average diameter of 0.2 mm

Perthite and microcline, in subhedral to anhedral crystals, comprise 36.5 per cent of this specimen, and twinned oligoclase (An_{12-14}), also subhedral to anhedral, makes up 23 per cent. The quartz grains, somewhat larger than the feldspars, are exceedingly irregular in outline and all are severely strained. Biotite, pleochroic in deep olive-green and yellow-brown, is scattered evenly throughout the rock. Primary magnetite is an inconspicuous accessory.

The minerals of this rock are slightly weathered. Perthite and microcline show meager kaolinization and sericitization, but oligoclase is conspicuously more altered and generally clouded with specks of kaolinite and flakes of sericite. The sericite ranges from tiny specks to large flakes 0.4 mm. in diameter, with an average diameter of 0.1 mm. It comprises 1.6 per cent of the granite. Some biotite has changed to chlorite with the concurrent separation of magnetite along cleavages, and some of this magnetite has weathered to hematite.

Slightly weathered granite (specimen 7). This specimen is from an outer shell and 6 to 8 inches outward radially from specimen 6. Megascopically it resembles closely specimen 6 except that it is slightly deeper pink as a result of more intense weathering. In thin section (fig. 11), specimen 7 is texturally nearly identical with specimen 6 but its mode differs somewhat. Perthite and microcline are greater by about 5 per cent but quartz is nearly 6 per cent less. The modes may be compared in Table III.

The most significant difference between this specimen and the preceding one lies in the intensity of alteration of the oligoclase. Nearly all grains are deeply clouded with kaolinite so that their twin lamellae are generally obscured, and flakes of sericite are both more abundant and larger. Reference to table III indicates that there is twice as much sericite in this rock as in specimen 6. This mineral ranges from tiny specks to flakes 0.6 mm. across, with an average diameter of 0.2 mm.

SIGNIFICANCE OF MICROSCOPIC OBSERVATIONS

Microscopic study of the spheroidal boulders described above leads to the following general conclusions regarding their development:

1. In general the cores of the spheroidally weathered boul-

ders consist of relatively fresh rock, but the shells surrounding the cores are more highly weathered. The intensity of this alteration increases outward more or less regularly through the successive shells. In the porphyritic olivine basalt, the interior boulder is fresh, but the inner shells are slightly weathered and the outer ones intensely so. The major alteration has occurred in the groundmass, the augite granules having partly oxidized to limonite, and the labradorite laths weathered to montmorillonite. Iddingsite rims around the olivine phenocrysts have also oxidized to limonite. Weathering of the vesicular basalt is even more marked. As one progresses from the fresh core outward, the augite crystals lose their birefringence, become clouded with specks of limonite, and locally seem to have altered to serpentine. Labradorite first shows alteration to sericite, kaolinite, and epidote on the periphery of the interior spheroid, and is finally converted to a fine aggregate of montmorillonite in the outer shells. Alteration of the biotite granite is less intense, but it is clear that the outer rims are more weathered than the inner ones. The chief changes are kaolinization and sericitization of the oligoclase. In the outer shells sericite is twice as abundant as in the inner ones and the flakes average twice the size.

2. In the weathering process, oxygen, water, and possibly some carbon dioxide were added to the rocks. New minerals, namely kaolinite, sericite, montmorillonite, serpentine, chlorite, hematite, and limonite were formed, most of these having specific gravities that are lower than those of the original minerals from which they altered. The result of these changes appears to have been an increase in the volume of the altered rock, the outermost shells suffering a more marked increase than the inner ones (fig. 4). Positive evidence of this increase is furnished by the outer layers of vesicular basalt in which the walls of the vesicles were apparently crowded inward by the increasing pressure. The amount of volume increase that took place in any of these rocks cannot be computed accurately, since the amount of material removed in solution is not known. Larsen (1948) found that the gruss surrounding spheroidally weathered boulders in Southern California, "has had some of its iron oxidized, has gained some water and much SiO_2 [leached from above], has lost a little K_2O, and gained a little Na_2O." (p. 119). Although these differences are small,

he concluded that they were adequate to expand the rock and cause it to crumble. He states: "A slight hydration of biotite and other minerals is probably sufficient to effect the change in volume that produces the disintegration and formation of the boulders." (p. 115). Leith and Mead (1915, p. 5-19) calculated that the weathering of a Georgia granite had caused a volume increase of 51 per cent, but since this rock was deeply decomposed, its volume increase must have been unusually large. The rocks discussed in the present paper probably suffered a much smaller increase, even in the outermost shells.

3. It appears probable that the spheroidal scaling of these rocks resulted from the oxidation and hydration of silicate minerals (fig. 4). In each case the fresh rock was first subdivided by joint planes into rectangular blocks, thus enabling subsurface water to attack each block from all sides. Weathering or alteration ensued along all faces of a block, and especially at edges and corners where the specific surface was greater. As a result, swelling of the outer surface of the block took place and internal stresses were set up. When these stresses had reached sufficient intensity, the rock split along a curved surface at a certain depth within the block where the cohesive strength of adjacent grains was exceeded. A concentric shell was thus produced. Further weathering along this new fracture produced additional shells within.

SPHEROIDAL BOULDERS WITHOUT SHELLS

Upon weathering some igneous rocks crumble to grains rather than scale off in shells, and the solid interior spheroid is bounded externally by a gruss of mineral fragments rather than by concentric shells. This is also spheroidal weathering. Whether a rock splits into shells or crumbles into grains appears to depend upon the cohesive strength between the mineral constituents of the rock. If this strength is great, adjacent grains will cohere, and, instead of crumbling, the rock will split along a curved surface at a certain depth within the boulder where the cohesive strength of the grains is exceeded by the shearing stress set up in the expanding outer portion. On the other hand, if cohesion is weak, due either to alteration along crystal boundaries or to the non-interlocking character of adjacent grains, the rock will crumble rather than separate as distinct shells. Fine-grained tough rocks, such as basalts and diabases which have well-developed ophitic texture, generally have sufficient cohesion between grains to develop distinct shells upon weathering. Coarse-grained granites and basic rocks, however, whose grains are less firmly interlocked, generally crumble to a coarse angular gruss surrounding a spheroidal core.

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REVIEWS

Introduction to Historical Geology; by RAYMOND C. MOORE. Pp. 582, 364 figs. New York, 1949 (McGraw-Hill Book Co., Inc., $5.00).—This new textbook by Professor Moore, of the University of Kansas, is written for the elementary student who has had not more than one semester of geology. The first chapter is very helpful in that it brings together many isolated facts covered by the ordinary course in physical geology and shows how they are used in unraveling the past record of our planet. The problems of correlation, the geologic time scale, and the length of geologic time are also included A thorough understanding of the contents of this chapter should do a lot towards preparing the student to appreciate the "Historical" approach to geology. The second chapter gives the background of evolutionary theory that is needed by one who has not studied biology. Chapter Three deals with the organization of the Solar system and a discussion of hypotheses that have been propounded to account for the origin of our planet. Beginning with Chapter Four, "The Pre-Cambrian Era," the book turns to a consideration of the interpretation of geologic history as shown by the rocks. This chapter and twelve that follow cover the usual material found in texts, to wit: distribution and character of formations, tectonics, vulcanism, economic products, and a little on the life of a particular period.

Chapter Twelve, "Nature and Evolution of Paleozoic Life," follows the chapter on the Permian and brings together in one place a mass of information, which in some books is scattered throughout the text. The wisdom of the plan used in this book may be questioned as students might wait until seven geologic periods have been studied before they come to a detailed description of some of the forms that lived during Cambrian time. Granted that it may be possible to give a more coherent description of Paleozoic life in a single chapter, nevertheless the reviewer wonders whether the student will get a clear picture of the life of each individual period. A possible solution to this problem would be to place this particular chapter ahead of the one on the Cambrian. A brief mention of the various forms of life would then be all that was necessary in each of the chapters on the Paleozoic. Similar chapters on Mesozoic and Cenozoic life follow the chapters on the Cretaceous and Quaternary. Changing their position in the text might also be considered. Chapter Twenty deals with "The Geologic Record of Man." An appendix of 28 pages, "Characters of Some Animal Groups Represented Among Fossils," closes the book This appendix, which starts with the Protozoa and ends with the Amphibia, duplicates much that has been covered by the previous chapters on the life of the various eras and contains some information that might well have been included earlier in the book. Thus a student studying the Ordovician, where graptolites are mentioned for the first time, must look them up in two other parts of the book to find out what they are like. Reorganization and consolidation of material might eliminate the need for the appendix.

As so often happens in first editions, there are numerous errors or doubtful statements that appear in the book. Some examples are given below. The letters under the illustrations of the fossils in Figure 139 do not agree with the descriptive text. On pages 149, 152, and 278, the author has placed the Eurypterids in the Crustaceans, although on page 511 they are correctly classified as Arachnida. Scorpions and spiders are also erroneously listed under Crustaceans on page 278 and correctly classified on page 511. If this was done in an attempt to simplify terminology, it is a dangerous policy The statement (pages 127-129) that in the Ordovician "of the Appalachian geosyncline and the interior platform as far west as the Mississippi" the bentonite beds are "Commonly . . . one or two feet thick" gives a false impression, as a great percentage of these beds are measured in inches, not feet. An example of a sentence that can cause trouble for students is found on page 69. "The unfossiliferous Pre-Cambrian rocks differ broadly from Cambrian and younger formations in physical constitution and structure." No matter whether this is applied to igneous, sedimentary, or metamorphic rocks, it does not give the correct impression, for younger rocks of all three types can be cited that have been or could be confused with Pre-Cambrian rocks. There is nothing in physical constitution or structure that is typically Pre-Cambrian. Although most of the illustrations are excellent, there are a few that need to be replaced, as for instance, Figure 59, the title of which states that it "shows cliffs of Paleozoic strata lying nonconformably on tilted Pre-Cambrian formations." One can study this picture for a considerable time and still not be sure where the tilted Pre-Cambrian strata are to be found in it.

It is hoped that when a second edition of this book is printed most of the questionable portions will have been changed.

LAWRENCE WHITCOMB

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FABRIC CHANGES IN YULE MARBLE AFTER DEFORMATION IN COMPRESSION

ELEANORA BLISS KNOPF

PART 1

ABSTRACT. Experimental deformation of a monomineral aggregate that is either statistically isotropic or statistically simple in fabric, carried out under controlled pressure-temperature conditions should give information on the process of deformation, provided that the fabric is studied before and after deformation by petrofabric analysis

Yule calcite marble from Colorado has a relatively simple pattern of preferred orientation both dimensional and crystallographic and was therefore used by David T. Griggs for deformation in compression under confining pressures of about 10,000 atmospheres, at 20°C and 150°C, both dry and in the presence of carbonated water vapor. Petrofabric analysis proved that the deformation was essentially homogeneous.

In the undeformed marble, calcite was arranged in a distinct planar structure (*s*) = (*ab*) oriented parallel to the north-south vertical plane. Petrofabric study shows that this planar structure is determined by triaxial ellipsoidal grains of calcite oriented with the shortest axis of the ellipsoid normal to *s*, which is the "grain" of the quarry structure. The optic axes of the calcite grains tend to lie nearly normal to *s* but make an angle of about 26° with the pole of *s*, which is the east-west direction in nature.

Deformation was produced in three sets of cylinders, oriented so that the cylinder axis in one set was normal to *s*, in the other two sets, parallel to *s* but mutually perpendicular. In dry deformation at 20°C the maximum strength was in cylinders whose axis was normal to *s* The stress-plan in these cylinders is unfavorable for twin-gliding because the orientation pattern of the undeformed marble would throw the direction sense in twin-gliding against the applied force in compression thus precluding movement by twinning The preferred orientation of the calcite lattice would allow translation-gliding to operate in these cylinders in conformity with the Law of Maximum Resolved Shearing Stress. Deformation is an axial strain and the cross section of the deformed cylinder is circular In a cylinder whose axis was parallel to *s* the strain was orthorhombic and the cross section after deformation was elliptical with the major axis of the ellipse always normal to *s*. This orientation would be expected from the original preferred orientation of the calcite in the undeformed fabric.

In dry deformation where the cylinder axes are parallel to *s*, the fabric of the undeformed marble is drastically rearranged into a new planar structure (*s*) = (a_2b_2) that is normal to the compressive force, and the optic axes of the calcite grains have assumed a new position nearly paral-

lel to the compressive force. The glide lamellae are close set and in many places bent and twisted. Many grains are cloudy and turbid, so that the microscopic texture shows no notable evidence of recrystallization even when the deformation was carried on at 150°C.

In deformation at 150°C in the presence of carbonated water, the fabric is entirely different. The grains are clear, lamellae are sharp and straight. The texture is clearly that of a recrystallized fabric. Unfortunately, reliable conclusions about fabric reorganization cannot be drawn from the single thin section that was cut from this cylinder, because of doubt as to the orientation of the section.

The present study of deformed Yule marble indicates the operation of intragranular gliding in conformity with the Law of Maximum Resolved Shearing Stress and suggests intragranular movement on two sets of intersecting glide planes. Conclusive proof of this mechanism, of the relative influence of twin-gliding and of translation-gliding, and of the crystallographic nature of operative glide systems will require further experimental work accompanied by detailed measurements of calcite axes and lamellae, grain by grain, in sections that are thinner than those used in the present study.

The present study has established that a radical change in preferred orientation, both dimensional and crystallographic, is caused by the deforming movement and that this change is definitely related to and controlled by the fabric pattern of the undeformed marble

INTRODUCTION AND ACKNOWLEDGMENTS

A SERIES of marble cylinders, whose internal fabric had been studied by petrofabric analysis, was deformed in compression by Mr. David T. Griggs in the Harvard Physics Laboratories in 1936. In this paper is recorded the result of petrofabric analysis of these cylinders both before and after deformation. Continuation of the program for rock deformation as originally planned was interrupted by World War II. In 1943[1] an advance report on the work completed before the war was given before the American Geophysical Union. The present paper describes the preliminary work in detail and outlines suggestions for further work. The manuscript has been read by Professor D. T. Griggs, who has very kindly furnished the data on the experimental deformation, and by Professor F. J. Turner of the University of California, to both of whom I tender sincere thanks, as well as to Professor Horace Winchell of Yale University, with whom I have discussed numerous problems. Dr. James F. Bell kindly furnished the diagrams showing fields that are susceptible and non-susceptible to twin-gliding and the photograph of calcite

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[1] Knopf, Eleanora Bliss. Fabric changes induced by Experimental Deformation of Marble. Am. Geophysical Union Trans. Pt. I, 271-272, 1943.

deformed by translation-gliding. My warm thanks are due to Professor George W. Bain of Amherst College, consulting geologist for the Vermont Marble Co. for data on the physical characteristics of the Yule marble and to the officials of the Vermont Marble Company who kindly facilitated my visit to the Yule quarry in Colorado for the purpose of collecting test material. The cost of cutting oriented cylinders for the deformation tests was covered by a grant-in-aid from the National Research Council and the experimental work was supported by the Committee on Geophysical Research and the Society of Fellows of Harvard University. Part of the apparatus for deformation at high confining pressure was constructed under a grant-in-aid from the Penrose Research Fund of the Geological Society of America. It is a great pleasure to acknowledge the kind assistance of these four organizations.

SCOPE OF THE INVESTIGATION

It is well known to metallographers that after a polycrystalline metal has been rolled or drawn it acquires a "texture", which is characteristically different for different kinds and degrees of working. By texture the metallographer means a concentration of certain crystal planes and vectors in definite direction with reference to the plane of working.

Many years before the metallographer recognized metal "textures" the geologist had discovered that deformed rocks have a "preferred orientation", in other words, a statistical concentration of crystal elements or of dimensional vectors in a definite direction with reference to the megascopic structure of the rock. In 1930 Sander[2] published numerous petrofabric diagrams illustrating the preferred orientation of various minerals in rocks of various sorts. In this way he was able to show that certain distinctive patterns of preferred orientation characterize rocks of different types. In the succeeding decade many other petrofabric diagrams have furnished a vast amount of data. The emphasis in all this work was laid upon the *kinematics* of the deforming process, which could be recognized by petrofabric analysis. As noted by Durand, writing in 1928 with the keen perception characteristic of so many French petrographers, "the dynamics is a much more delicate matter", which cannot be established conclusively by *this* method of studying deformed rocks.

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[2] Sander, Bruno. Gefügekunde der Gesteine. J. Springer. Wien, 1930.

The mechanism of the orienting process in rock minerals has been largely a matter of inference, because the data hitherto obtained by petrofabric analysis represent the end result of the process, and up to 1936 only two attempts had been made to relate preferred orientation in rock minerals to experimentally induced deformation; and in both of these experiments the applied force produced ruptural and not plastic[3] deformation.[4] In metal working, on the other hand, the process has always been a matter of observation, because the resultant texture is actually produced under controlled conditions.

In this way it has been established that the change in shape in deformed metal is accomplished by intragranular slip, the movement starting in those glide planes[5] on which the resolved shear stress in the glide direction is a maximum. As slip proceeds, the initially operative glide planes change position with reference to the direction of applied force and in doing so bring other glide planes, originally less favorably situated, into a position where they can operate to carry on the deformation. Intergranular movement is probably of subordinate importance in developing a preferred orientation in polycrystalline metal fabrics. The ability to deform by translation-gliding is more quickly attained in those metal crystal lattices that have several possible translation systems, such as in the cubic lattice of aluminum, than it is in the hexagonal lattice where the possibility of movement is restricted to one crystal lattice plane as in zinc. In zinc the ability to twin on a different lattice plane from the translation plane facilitates deformation.

Petrofabric analysis of deformed rocks (tectonites) had indicated the probable influence of the same mechanism of intragranular slip in rock deformation previously recognized in metal deformation. In addition to the slip mechanism, it had also suggested the influence of grain rupture, followed by a dimensional rearrangement of particles into a preferred orientation governed by grain shape. When recrystallization increases the grain size, the preferred orientation of these seed particles is retained.

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[3] Plastic is used here to characterize deformation exceeding the elastic limit in which the spatial continuity of the material is unimpaired

[4] Sander, B, Felkel, E., and Drescher, F. K Festigkeit und Gefugeregel am Beispiele eines Marmors. Neues Jb, BB59A, 1-26, 1929

Bell, J. F. Festigkeit und Gefügeregel am Beispiel eines Granits. Neues Jb. BB71A, 193-213, 1936

[5] "Gliding" comprises displacement by translation and by twinning Slip and gliding are used here synonymously

In 1934 it occurred to the author that the experiments on rock deformation under high confining pressure that were being conducted by Mr. David T. Griggs at Harvard University would furnish an ideal field for petrofabric investigation of experimentally deformed rocks. These experiments were being carried out under accurately controlled pressure-temperature conditions and therefore should give definite information on the nature of the orienting process, provided that the fabric of the rock is studied both before and after deformation.

An ideal rock fabric to be tested, under a given stress set-up, should be a statistically isotropic, homogeneous, monomineral aggregate, of grain size sufficiently large to be readily studied with the universal stage in measuring the orientation of the individual grains, and not so large but that in one thin section enough grains can be measured to give statistically a true sample of the whole.

In such an ideal rock fabric the mineral grains have a random orientation (a statistical scattering of position), for as soon as the individual grains show a definite alinement, either dimensional or crystallographic, the fabric is anisotropic.

SELECTION OF MATERIAL FOR STUDY

As Griggs had run a number of successful experiments on the deformation of limestone and marble at high confining pressure,[6] a calcite rock seemed a desirable monomineral aggregate to use in fabric studies. Experimental study of single calcite crystals had proved that the mechanism of deformation is by mechanical twinning and had suggested the concomitant influence of translation-gliding. The gliding planes are three rhombohedral planes; therefore considerable flexibility of deformation is expectable. It soon became apparent that in order to obtain the necessary grain size the material chosen would have to be a metamorphic marble (a calcite tectonite),[7] because homogeneous calcite marbles of depositional original could not be found.

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[6] Griggs, David T Deformation of rocks under high confining pressures. Jour Geology, 44, 541-577, 1936.

It appeared possible that calcite grains in a limestone that had been marmarized by contact-metamorphic action would prove by statistical study to be of random orientation, provided that the thermal action had not been accompanied by a penetrative movement that would cause a preferred orientation of the constituents. The marble from the Yule quarry in Colorado has been considered to be a thermally altered limestone of late Paleozoic (Mississippian) age[8, 9] and, unlike many contact-metamorphosed marbles it lacks contact silicate minerals. Therefore, in 1934 the author visited the quarry of the Vermont Marble Company from which came the block of "Golden Yule" marble used in the tomb of the Unknown Soldier in the National Cemetery at Arlington, Virginia. This quarry is on the west side of Yule Creek, three miles south of Marble, Gunnison County, Colorado.

The marble at the base of the quarry about 200 feet below the surface entrance appeared to be ideally suited for the purpose, so far as field observation could indicate It is a uniform and approximately even-grained white marble and the only megascopic structure evident in the quarry is a set of fractures or joints running N65°E and dipping about 80°NW. Another set of discontinuous joints striking N55°W and dipping about 70°SW form *en échelon* zones not easily seen. The absence of bedding planes makes it possible to saw the marble into exceptionally large blocks. However, in spite of the massive character, the quarrymen recognize a certain anisotropy in the rock and they report that the "grain" of the marble,[10] which is the plane of easiest splitting, strikes approximately north and dips westward at almost 90°. (See pl. 1).

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[7] A tectonite is a rock in which deformation has been accomplished by interrelated movements of individual fabric elements in such a way that the individual movements can be integrated into the movement as a whole. In a non-tectonite individual movements of components are independent and unrelated. Therefore, they imply nothing about movements of other components.

[8] Vanderwilt, J. W, and Fuller, H. C Correlation of Colorado Yule marble and other early Paleozoic formations on Yule Creek, Gunnison County, Colorado. Colo. Soc. Proc., 13, 7, 439-464, 1935.

[9] Vanderwilt, J. W. Geology and Mineral Deposits of the Snowmass Mountain area, Gunnison County, Colorado. U. S. Geol. Survey Bull. 884, 19-21, 1937.

A specimen of marble, measuring more than one cubic foot, was collected from near the quarry floor and the geographic orientation was marked on the block together with the direction of the grain.

In the laboratory a thin-section block was sawed from this large specimen with the orientation indicated in figure 1, a.

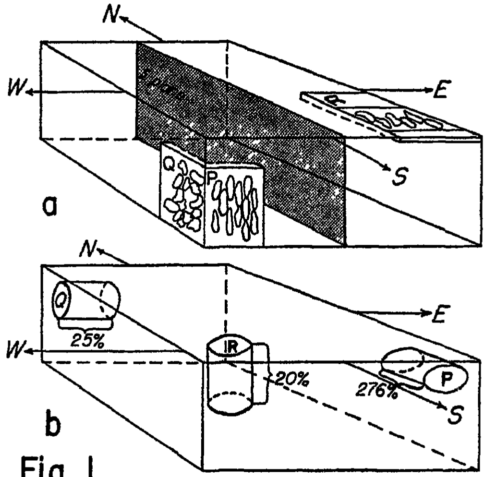

Fig. I

Figure 1 Orientation of thin sections and test cylinders of Yule marble before deformation in compression

a) Thin section block with 3 mutually perpendicular sections P Q.R. Megascopic *s* plane is determined by dimensional orientation of discoid grains

b) Orientation of test cylinders. Numbers indicate percentage reduction of original cylinder after deformation Dimensions of original cylinder, 1 inch x ½ inch.

The faces were lettered as follows:

1. P, vertical surface; normal to strike and normal to the dip of the grain.
2. R, horizontal surface; parallel to strike and normal to dip of grain.
3. Q, vertical surface; parallel to strike and dip of the grain.

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[10] In marble quarrying "grain" is used as synonymous with "rift" in granite quarrying.

COMPOSITION OF THE YULE MARBLE

A small chip of the rock, about 2 cc. in volume, dissolved almost completely in cold dilute HCl within a few hours. A surface of about 3 cm. x 4 cm. was polished and stained with a molar solution of $Cu(NO_3)_2$ for less than three hours.[11] On fixation with NH_4OH the whole surface assumed a uniform blue color. These two tests indicate that dolomite is absent and that the Yule marble is practically a pure calcite marble.

The following analysis shows that the rock is made up almost entirely of calcite.

Chemical analyses of clear Yule Colorado Marble[12] *made under direction of G. W. Smith, Case School of Applied Science, Cleveland, Ohio, Oct.* 22, 1907.

$CaCO_3$	99.50
$MgCO_3$	.19
$FeCO_3$	.03
$MnCO_3$	.02
SiO_2	.05
Al_2O_3	.03
Fe_2O_3	trace
MnO_2	none
$CaSO_4$	.09
Undet	.09
	100.00

PHYSICAL CHARACTERISTICS OF THE YULE MARBLE

Bulk specific gravity is 2.70. True specific gravity is given by Griffiths[13] as 2.72. Porosity is low; about 0.15 per cent and elasticity is high. (Young's modulus is 4.01×10^5 kg/cm^2 when the force is applied parallel to the veining and 3.84×10^5 kg/cm^2 when the force is applied normal to the veining).[14] Griffiths[15] reports a porosity of 0.45 per cent and absorption of 0.19 on a specimen of Yule marble. When the

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[11] Rodgers, John Distinction between calcite and dolomite on polished surfaces Am. Jour. Sci., 238, 797-798, 1940.

[12] Vanderwilt, J. W. Geology and Mineral Deposits of the Snowmass Mountain Area, Gunnison County, Colorado. U. S. Geol. Survey Bull. 884, 160, 1937.

[13] Griffiths, J. H. Physical properties of typical American rocks. Iowa Engineering Experiment Sta. Bull. 131, 12, 1937.

[14] Data furnished by courtesy of Dr. G. W. Bain, Consulting Geologist of Vermont Marble Co.

[15] Griffiths, J. H., loc. cit.

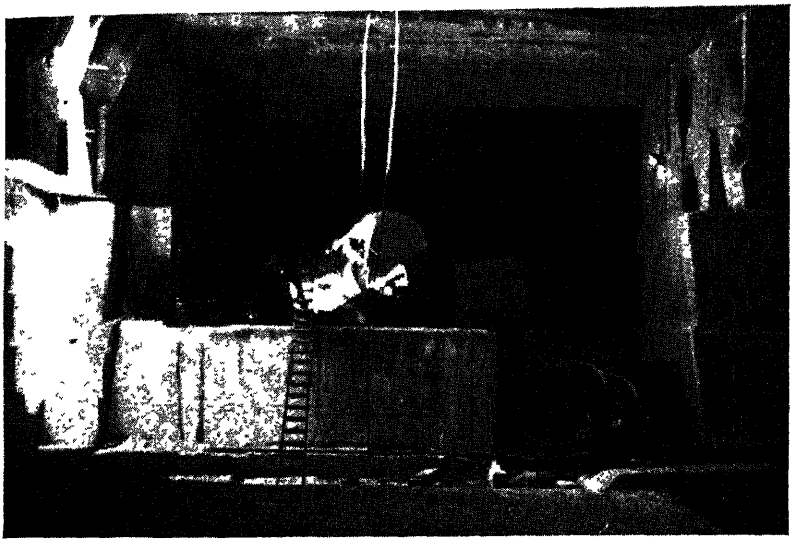

Plate 1 West wall at the floor of Yule Quarry, Colorado Looking northwest Shows method of sawing large blocks, dip of the "grain" in this face is steep towards the southwest (to the left) Courtesy G W. Bain

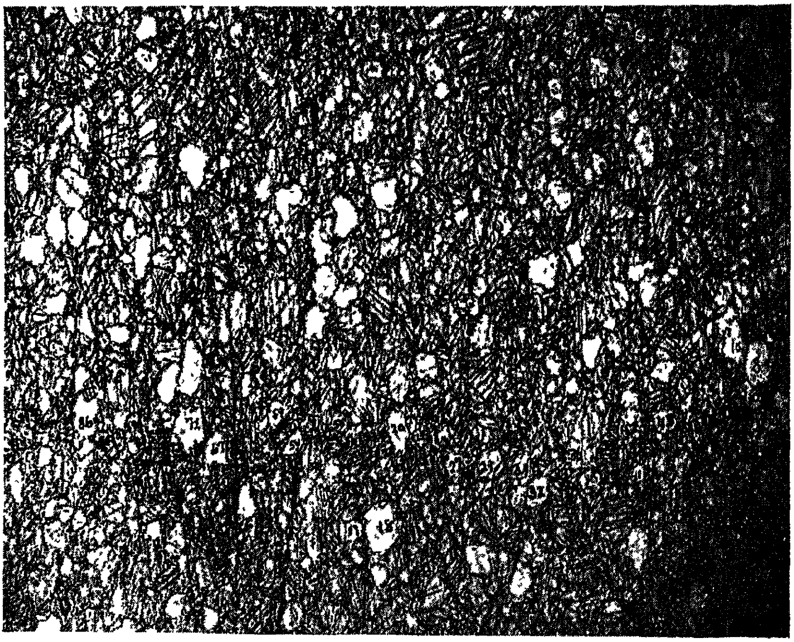

Plate 2 Thin section P_1 cut from the P face of the thin section block, x 10. Numbered grains represent traverses across the thin section

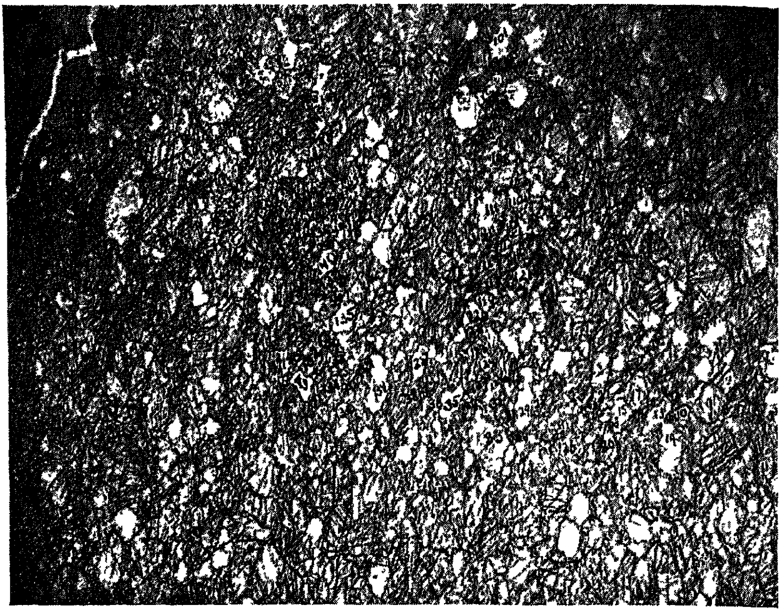

Plate 3. Thin section R_2 cut from the R face of the thin section block, x 10

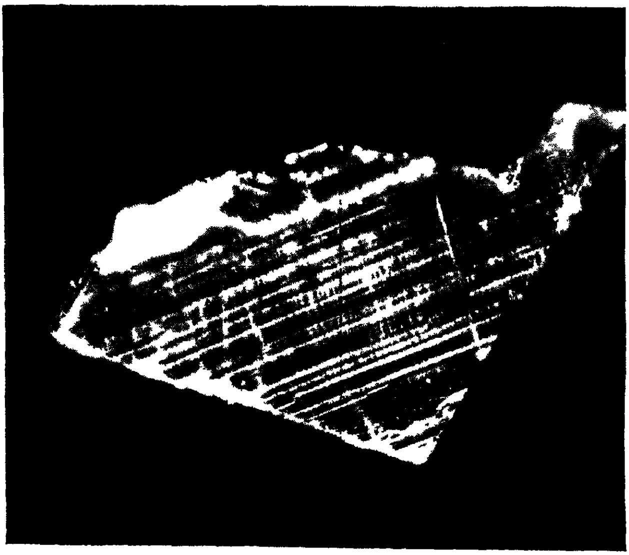

Plate 4 Translation-gliding striae in calcite crystal induced by a compressive force applied approximately parallel to the optic axis of the crystal In this position twinning could not take place as the shearing stress on the $(01\bar{1}2)$ glide planes would be resolved in the wrong direction sense. Courtesy of J. F. Bell.

force is applied normal to the veining there is no hysteresis. On relief of force applied parallel to the veining, hysteresis amounts to 0.040 per cent. Permeability is moderately high (it requires 71 minutes for water to penetrate the stone to 1.5″). In one test run by Griggs and Bell on an R cylinder the crushing strength at room temperature parallel to the "grain" was 4220 psi = 296 kg/cm². Bain reports that in a direction normal to the "grain" the strength could be 8000 psi. In 1948 Professor Henry A. Lepper, Jr. of the Engi-

Linear Thermal Expansion of Yule Marble from 20°C. to Indicated Temperature, in Per Cent

(Rosenholtz, J. L., and Smith, D. T., Rensselaer Polytechnic Institute)

	1 Orientation E-W	2 Orientation N-S	3 Orientation—vertical
	‖ Maximum concentration of *c* axes of calcite.	⊥ Maximum concentration of *c* axes of calcite.	⊥ Maximum concentration of *c* axes of calcite.
	First Heating		
Temperature 0°C.	Extension in %		
100	0.14	0.03	0.03
200	0.53	0.19	0.17
300	0.92	0.36	0.34
400	1.28	0.51	0.52
500	1.70	0.74	0.72
600	2.13	0.94	0.94
700	2.52	1.14	1.14
	Elongation 1.02%	Elongation 0.50%	Elongation 0.71%
	Second Heating		
Temperature 0°C.	Extension in %		
100	0.05	—0.03	—0.021
200	0.16	—0.04	—0.033
300	0.31	—0.04	—0.027
400	0.42	—0.02	—0.005
500	0.62	+0.04	+0.06
600	0.96	+0.17	+0.20
700	1.25	+0.32	+0.37

neering School of Yale University tested strength to rupture in compression in two sets of oriented specimens of Yule marble.[16]

In the same year Professor Joseph L. Rosenholtz and Professor Dudley T. Smith of Rensselaer Polytechnic Institute determined the linear thermal expansion on Yule marble in three sets of oriented specimens. As this work is part of a recently begun, long-term program on determination of linear thermal expansion and transformation of crystallographically oriented minerals and rocks, Dr. Rosenholtz has very kindly released the data in the table (p. 441) for publication in the present paper.

The high values of the first heating cycle appear to be due to relief of some residual deformational strain, because the extension in each specimen at the end of the first heating-cooling cycle was a permanent elongation. On running the specimens through a second heating-cooling cycle, there was no further change in length at the end of the cycle.

PETROGRAPHIC FEATURES OF THE YULE MARBLE

The rock is a translucent, white, crystalline marble of medium grain. Careful inspection of both thin section and hand specimen shows that the calcite grains are distinctly elongate in the P and R face of the block, whereas in the Q face, which is cut parallel to the strike and dip of the "grain", they are practically equidimensional. In the P face the long dimension of the grains is parallel to the vertical axis in nature and in the R face the long dimension is in the north-south direction. (See fig. 1,a and pls. 2, 3.) Plates 2 and 3 are photomicrographs of thin sections P_2, and R_2 cut respectively from the P and R faces of the thin-section block. The orientation of these thin sections is shown in figure 1, a.

The average grain intercept is given in the following table:

	Average ratio between the longer and shorter grain intercepts measured in each section	Dimensions in Millimeters
P	18:1	9:.5
R	3:1	1.5:.5
Q	1:1	.5: 5

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[16] Lepper, H. A, Jr. Compression tests on oriented specimens of Yule marble: Am. Jour. Sci., vol. 247, August 1949 (in press).

This indicates an arrangement of ellipsoidal calcite individuals occupying the plane of the "grain" in such a way that the short axes of the ellipsoids are normal to the "grain" of the marble. The calcite forms a granoblastic fabric whose components, in general, are of nearly the same size. The grains are clear, sharply transparent, and show numerous lamellae separated by lines that become thin and sharp when turned on the universal stage into the position where the planes of separation are parallel to the microscope tube. Some grains show only one set of parallel lines. Many grains show two or more intersecting sets of parallel lines. In some grains it is difficult to decide by inspection whether these lines are the traces of twin-gliding planes or of cleavage planes or of translation-gliding planes. The positions of all the lamellae sets in each grain were measured and recorded on the projection net by plotting the pole of that particular set of lamellae planes. The way in which lamellae due to twin-gliding, translation-gliding, and cleavage were distinguished is described later.

PART I.

PRELIMINARY PETROFABRIC ANALYSIS OF YULE MARBLE (BEFORE EXPERIMENTAL DEFORMATION)

From the thin sections cut parallel to the P, Q, and R faces of the thin-section block petrofabric analyses of the calcite fabric were made by measuring and projecting on to the lower hemisphere of a reference sphere the positions in space of the calcite axes and of the poles of the calcite lamellae. These positions are all recorded on an equiareal (surface-true) net.

Measurement of lamellae. In measuring the pole of a given set of parallel lamellae the grain is rotated upon the microscope stage (in other words around the vertical axis) until the traces of the lamellae are parallel to one of the cross hairs.[17] The grain is then tilted upon the N-S (A_2) or the

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[17] The lamellae should be brought parallel to a cross hair in such a way that the vibration direction of the slow ray in the calcite is parallel to the vibration direction of the polarizer because then the index of refraction in the grain is the same as the index of refraction of the hemisphere. The hemispheres used in measuring calcite usually have an index of 1.649, which is ω in calcite. If the trace of the lamellae is placed so that the fast ray in the grain is parallel to the vibration direction in the lower nicol, the angular reading should be corrected for the difference in index between the grain and the hemisphere.

E-W (A_4) axis of the universal stage until the lines marking the intersections of the lamellae with the thin section are parallel to the microscope tube, in which position the lines show the maximum thinness and sharpness. However, lamellae that lie in or near to the plane of the thin section can never be turned up to a position parallel to the microscope tube owing to the fact that after the universal stage has been rotated for a certain angle (ϕ), about either the N-S (A_2) or E-W (A_4) axis, total reflection occurs. This angle (ϕ) depends upon various factors such as the relative indices of the hemispheres and of the mineral grain, the microscope constants, and so forth. In the present work total reflection occurs after a rotation of about 45°. This gives a circular "blind spot" or "unmeasurable area" on the projection net with a radius equivalent to a distance of 45° from the center of the net. Whenever two mutually perpendicular sections of the material are available, it is possible to fill this "blind spot", by utilizing a second section at right angles to the first to measure enough poles that the area common to both projections contains an equal number of poles. Then these poles from the second section occuring in the area that corresponds to the blind spot of the first section are themselves rotated to the position of the first section and thus the unmeasurable area is filled,[18] and a complete diagram is prepared.

The completeness with which the blind spot in any diagram can be filled depends upon the angle at which total reflection occurs—if the angle of total reflection (ϕ) is $> 45°$, the radius of the corresponding blind spot must be less than 45° and therefore the blind spot in one section must lie within the measurable area in the section at right angles to the first. If the radius of the blind spot is > 45 ($\phi < 45°$) a part of the blind spot of the first section will lie within the unmeasurable area of the second section and some poles will always be missed in both sections. As long as the radius of the blind spot is not greater than 50° any essential concentration that is missed in two sections that are mutually perpendicular, for example P and R, can be found in the third section, Q, normal to the other two.

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[18] Knopf, E. B., and Ingerson, E. Structural Petrology. Geol. Soc. America Mem. 6, 230-233, 1938.

COMPLETE DIAGRAMS

As the contoured diagram represents relative density of measured poles, it is obvious that a petrofabric analysis can never represent the measured field correctly until the unmeasurable area is filled. Therefore, it is absolutely useless to draw final conclusions about significance of maxima, presence or absence of girdles, etc., until the diagram is completed. However, if the diagram shows a strong peripheral concentration of poles, it is advisable to contour the incomplete diagram in order to evaluate the maxima that are peculiar to the diagram that has been measured, some of which will not reappear in the measured area of other sections. For example, suppose we measure a fabric in which there is a strong concentration of lamellae poles around the emergence of the east-west axis causing a maximum around the east-west pole of a *horizontal* section. If we measure a diagram from a section *at right angles to the first and looking north* the maximum recorded in the first analysis will reappear in the second diagram. But suppose we measure a section *at right angles to the first and looking east or west* the maximum around east or west will never be found. Its absence in the third section is no evidence of inhomogeneity in fabric and similarly its presence in the third section when filled from either of the other two is no evidence of homogeneity because that maximum can never be recognized by measurement anywhere except in the sections that are parallel to the east-west axis.

Moreover the fact that we have in the center of the calcite lamellae diagrams a blind spot with a radius of about 45° means that in any one section there is outside this blind spot a peripheral area where the density is indeterminate because the relative density of concentration of the measured poles in the given section will be affected by the poles that lie in the unmeasurable area and thus do not show in the measured section. Within the periphery of the projection net there is a zone where the number of poles in the given counter circle is not changed by any poles that are outside of this zone. Likewise in the blind spot itself there are no poles within the counter zone. But in the transition zone between the peripheral zone of the projection net and the circumference of the blind spot, a counter will catch the poles that were actually measured in the measurable area but will also include a part of the unmeasurable area which cannot be included in the count until the blind spot is filled by rotation of poles from another section. When the diagram is completed a submaximum that showed up in counting the diagram measured may shift its place or even "wash out" completely. Therefore, while there is a certain value in contouring an unfilled diagram it should *never* be used for final conclusions without also contouring the completed diagram. What appears to be a good single peripheral girdle in one measured diagram may prove to be two or even three girdles when the diagram is completed, a difference that would have an important tectonic significance in a regional study.

SELECTIVE DIAGRAMS

In order to facilitate the study of selective "elemental diagrams", i.e., diagrams prepared by grouping the measured data from certain kinds of grains, selected with reference to grain size, presence or absence of lamellae, etc., the thin sections should be photographed and the grains in the photomicrograph should be numbered consecutively. Then each grain measured can be identified upon the photograph and measured data, lamellae poles or optic axes, edges between lamellae, etc., can be recorded against the appropriate number. Such a record makes it easy to construct diagrams from any desired grouping of grains.[19]

HOMOGENEITY OF THE FABRIC

If a statistical analysis of preferred orientation of grains in a given fabric is to be used to draw sound conclusions about the behavior of the fabric under the application of a deforming force, the field of matter under consideration must be statistically homogeneous. In other words, whatever may be the structure of the field under consideration, whether it is statistically isotropic or anisotropic, any two parts that are similarly oriented must be identical.[20]

[19] In early stages of the work the importance of correlated grain measurements was not appreciated, but all diagrams made later were prepared in this way.

[20] The importance of determining homogeneity in petrofabric analysis of a rock, which seems to have been overlooked recently by some workers, is so great that the procedure is described here in detail.

An easy way to test homogeneity in the *area* technically known as the "field" of a thin section is to measure the grains in separate traverses that cross the thin section in different parts. The poles that make up the different traverses are contoured separately in elemental diagrams and are compared to determine whether any essential difference in fabric is indicated. All the measured poles are then collected into one "collective diagram", which is contoured and compared with the "elemental diagrams" of the individual, separate traverses. In a fabric that is statistically homogeneous all these diagrams, both collective and elemental, will show essentially the same pattern.

A fabric may be essentially homogeneous in the field represented by the *area* of the thin section but may be distinctly inhomogeneous in the field represented by the *volume* of a thin-section block. The test for this is made by comparing diagrams cut from different faces of the block. For example, a diagram is prepared from a thin section cut from the P face. Then it is geometrically rotated to the position of the R section by shifting the contours 90° along small circles that are normal to the axis common to both faces, which in the P and R faces is the east-west geographic axis. This *rotated* diagram shows the pattern as it would appear in a diagram *measured* from R if the fabric is homogeneous. The diagram actually measured in R is then compared with the diagram that was measured in P and rotated to R. If the diagrams do not show a reasonable correspondence it indicates an inhomogeneity in the field under consideration and by this comparison it is possible to determine the direction of shift in the fabric pattern of P in a third dimension at right angles to the area of the P section.

SIGNIFICANCE OF PATTERN

In any statistical study the data on which conclusions are based must represent a true sample of the whole that is being studied. For example one bad egg in a box of twelve eggs does not mean that in a crate of twelve dozen eggs there will be a bad egg in each dozen because one box may have chanced to get the only bad egg that was present in the whole 144 eggs. But if each of six boxes has one or two bad eggs it is much more likely that the ratio of bad to good in the whole crate will turn out to be 1:12 instead of 1:144. A great deal has been written about significance of maxima in petrofabric analysis and about the way to determine this significance mathematically. However, for purposes of fabric analysis the meaning of a given maximum depends actually upon its persistence and position within a given field of distribution. As it is impossible to evaluate homogeneity without making collective diagrams from numerous elemental diagrams, the easiest way to determine the significance of a given maximum is to watch it from one diagram to another during the course of the study. If the same maximum reappears in approximately the same place in a number of successive traverses in a thin section, or in a number of thin sections in a rock, the concentration of poles represented by the maximum is not fortuitous but is a significant indication of a preferred orientation that has a meaning in interpreting fabric. Moreover, if in several thin sections of a rock, the same relative arrangement of maxima recurs but in a notably different relation to the geographic coordinates, the shift in fabric pattern indicates a structurally significant shift in the fabric structure.

For that reason it is a good idea, when measuring several hundred grains to count first fifty, then one hundred, and so on in successive groups and to notice what parts of the pattern are constant from a count of a few poles up to a count of several hundred poles, also to notice which parts of the pattern disappear and which are intensified and perhaps shifted in successive counts.

The diagrams listed in the following table were measured and contoured by the author from calcite grains in the Yule marble. The orientation of the thin-sections from which the diagrams were measured is shown in figure 1, a.

To get an independent check on the diagrams from the P, R, and Q thin sections, measured by the author, 300 calcite lamellae in P_1, 100 in R_1, and 100 in Q_1, were measured and contoured by Dr. James Forbes Bell, who was at that time research assistant to the author. These diagrams checked in all essential particulars with the other diagrams from sections in the P, R, and Q faces.

In order to test the homogeneity of the whole thin-section block a second set of chips, labelled P_2, R_2, and Q_2, was cut from a different part of the thin-section block and the thin

Diagrams prepared from undeformed Yule marble.

From P face	From R face	From Q face
P_1 thin section	R_1 thin section	Q_1 thin section
Calcite lamellae		
90 poles (elemental) 150 poles (elemental) 305 poles (collective) 385 poles (complete collective filled from Q_1) 355 poles (complete collective filled from R_1) *150 poles*[21] *150 poles* *300 poles* (collective) 520 poles (collective)	48 poles (elemental) 72 poles (elemental) 120 poles (collective) 165 poles (collective for one traverse) 115 poles (from another traverse) 328 poles (collective from both traverses) 398 poles (complete collective from both traverses filled from Q_1) 403 poles (complete collective filled from P_1) *100 poles*	190 poles *100 poles* 375 poles (complete collective filled from P_1) 388 poles (complete collective filled from R_1)
Calcite optic axes		
40 axes (elemental) 75 axes (elemental) 210 axes (collective) 230 axes (collective)	20 axes (elemental) 69 axes (elemental) 109 axes (collective) 255 axes (collective)	
P_2 thin section	R_2 thin section	
Calcite lamellae		
165 poles (in one traverse) 165 poles (in another traverse) 330 poles (collective) 410 poles (complete collective filled from R_2)	165 poles (in one traverse) 166 poles (in another traverse) 331 poles (collective) 370 poles (complete collective filled from R_2)	
Calcite optic axes		
	38 axes	
282 axes collective from P_1 and P_2.		

sections P_2 and R_2, made from these chips, were measured for optic axes and lamellae in calcite. The elemental and collective diagrams from the P_1, R_1, and Q_1 faces respectively, and from the P_2 and R_2 faces respectively, checked well enough to

[21] Italics indicate diagrams measured by Dr. James Forbes Bell, at that time research assistant to the author.

indicate an essential homogeneity in the field of the thin section in both sets of thin sections. Moreover, the comparison of P_1 with P_2 and of R_1 with R_2 indicates that the fabric is essentially homogeneous within the field of the material from which the test cylinders were cut for deformation. For this reason no new Q diagrams were measured from the second set of thin sections.

PETROFABRIC ANALYSIS

OPTIC AXES

P diagrams. The analysis of the calcite optic axes diagrams shows a striking homogeneity of calcite fabric within the field of the thin section. A strong two-fold maximum Ax_1 and Ax_2 is clearly evident, striking about N65°W and separated upwards and downwards into two divisions. The north-south spread of this maximum is about 45°, the east-west spread about 30°. It only requires a count of 43 optic axes measured in the P_1 section (fig. 2) to indicate clearly the location of this two-fold maximum and the same maximum recurs in the count of 75 axes (fig. 3) and persists also in 210 axes all measured in P_1. This similarity in diagrams from all three counts shows the homogeneity of the field of the thin section. The diagram from 282 axes (fig. 4) measured in both P_1 and P_2 is practically identical with that from 210 axes measured in P_1, thus indicating that the fabric is homogeneous within the field of the thin section block. A submaximum (Ax_3) on the periphery of the diagrams, near the east-west pole, has begun to appear in the diagram of 75 axes and is definitely established in the collective diagram of 282 axes. The pattern in figure 4 is sharp,—a main maximum representing axes striking about N65°W with a tendency to swing upward and downward and a submaximum representing axes that lie horizontally in an east-west direction.

R diagrams. The optic axes from the R_1 section were measured with the same precautions to observe the homogeneity within the field of the thin section as in measuring optic axes from P_1 section. 122 optic axes measured from the R_2 section (fig. 5) showed such a good coincidence with 110 optic axes measured from R_1 (fig. 6) that a diagram was contoured from 242 axes measured in R_1 and R_2 (fig. 7). This represents the R field of the thin section block. In order to test

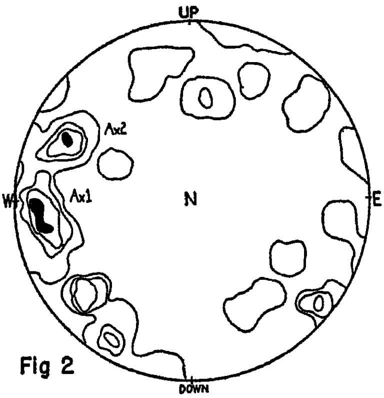

Fig 2

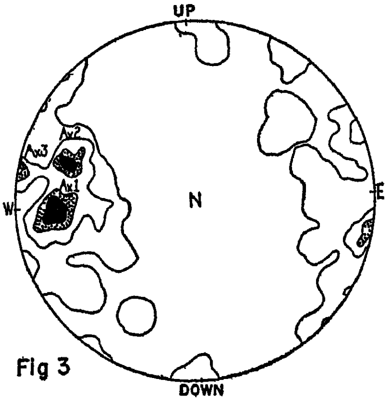

Fig 3

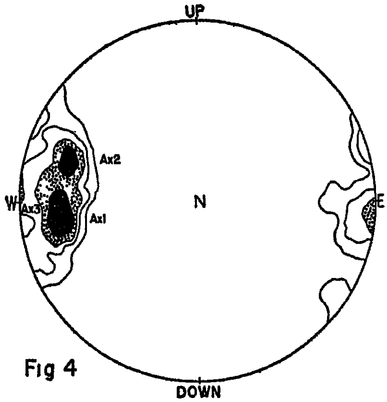

Fig 4

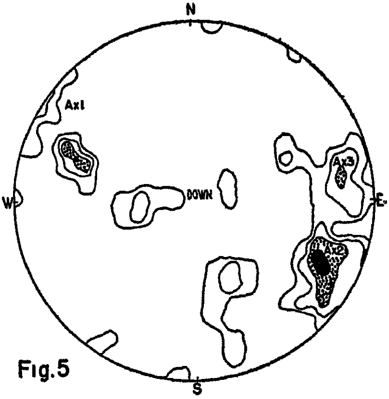

Fig.5

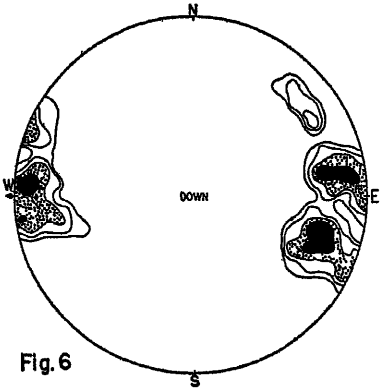

Fig.6

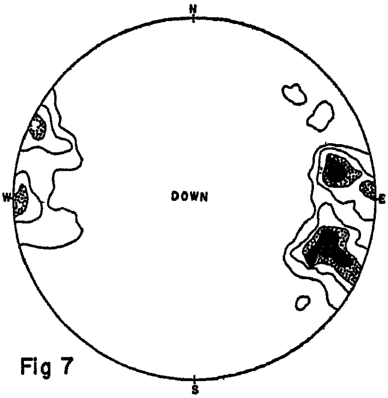

Fig 7

Figure 2 43 optic axes of calcite in P_1 section. Contours (10-9)-8-5-3-0%.
Figure 3. 75 optic axes of calcite in P_1 section Contours (12-9)-7-4-2-0%.
Figure 4. 282 optic axes of calcite from P_1 and P_2 sections. Contours (8-6)-4-3-2-0%.
Figure 5 122 optic axes of calcite in R_2 section. Contours > 6-4-3-2-0%.
Figure 6 110 optic axes of calcite iln R_1 section. Contours (8-6)-4-3-2-0%.
Figure 7. 242 optic axes of calcite from R_1 and R_2 sections Contours (8-6)-4-3-2-0%.

the homogeneity of the fabric in the field between P and R, this diagram of axes in R was checked against a diagram made by rotating the diagram of 282 axes measured in P_1 and P_2 to the position of R (fig. 8). The high degree of coincidence between these two confirms the conclusion that the fabric is essentially homogeneous within the field from which the test cylinders were cut for experimental deformation. The essential feature of the fabric is the concentration of axes into an approximately horizontal position striking about N65°W.

Q diagrams. No calcite axis diagram was made from Q thin section because this section is approximately normal to the concentration of optic axes. Therefore, the optic axes must be measured by setting them in the axis of the microscope tube, which is difficult to do with great exactitude in a section normal to the optic axis because it is impossible to determine the position of total darkness within a range of several degrees. In a mineral with the high birefringence of calcite, it is particularly difficult to determine the exact position of total extinction. Many grains in this section show a small and variable but distinct biaxiality. Therefore, as the essential homogeneity of the fabric has been established by rotating the lamellae and axes diagrams from P to R, it was decided the position of the axes in Q could be shown by rotating the axes measured in P_1 or R_1 to the position of Q as accurately as by actual measurement in Q (fig. 9).

LAMELLAE

P_1 and P_2 diagrams (unfilled collective). The diagram of 305 calcite lamellae from the P_1 face looking north (fig. 10) shows a clear segregation of poles into two 100° peripheral arcs extending about equal distances on either side of the east-west pole of the diagram. The vertical axis in Nature emerges in well-defined minima on the periphery of the projection net at 90° from the east-west pole. The same arcuate concentration of lamellae around the periphery reappears consistently in all the collective diagrams.

The collective diagram of 300 lamellae made by Dr. James F. Bell, shows a good correspondence with figure 10, which is the collective diagram of 305 lamellae measured by the writer.

Figures 11 and 12 are two elemental diagrams made from two traverses of 165 lamellae each, in the thin section P_2 which

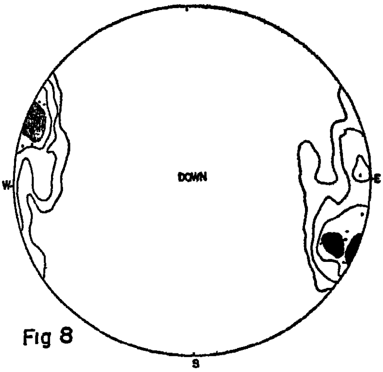

Fig 8

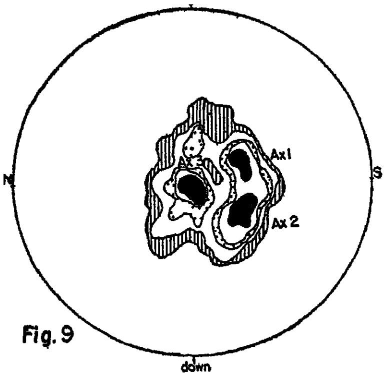

Fig. 9

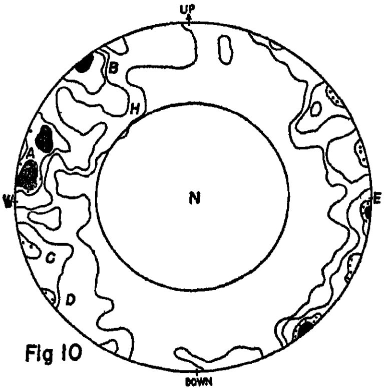

Fig 10

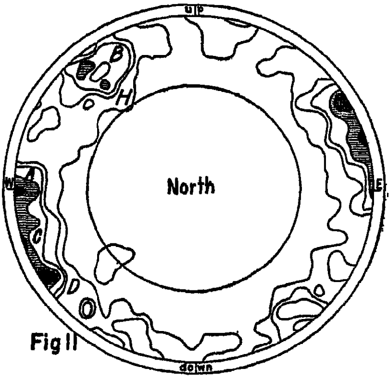

Fig 11

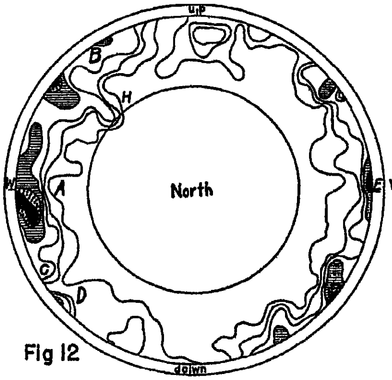

Fig 12

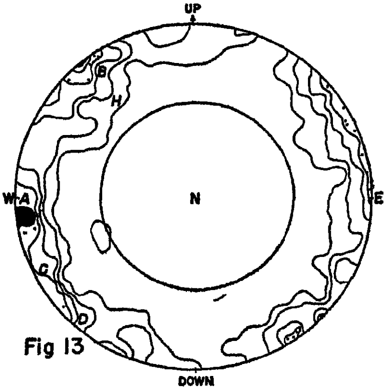

Fig 13

Figure 8 282 optic axes from P_1 and P_2 rotated to the position of R.

Figure 9. 160 optic axes of calcite measured in P and 160 measured in R rotated to the position of Q. Contours (7-6)-5-4-3-2-0%.

Figure 10. 305 poles of calcite lamellae in P_1 section. Contours (5-4)-3-2-1-0%.

Figure 11. 165 poles of calcite lamellae in P_2 section. First traverse. Contours (6-5)-4-3-2-1-0%.

Figure 12 165 poles of calcite lamellae in P_2 section. Second traverse. Contours (7-6)-5-4-3-2-1-0%.

Figure 13. Collective diagram of 330 poles of calcite lamellae from P_2 section. Contours (6-5)-4-3-2-1-0%.

was cut from a different part of the thin section block. A collective diagram of 330 lamellae in figure 13 was made by combining these two elemental diagrams from P_2. These P_2 diagrams are all essentially homogeneous with each other and with the P_1 diagrams. The peripheral maximum belt extends a little farther upwards from the emergence of the west pole in diagrams from P_1 than it does in diagrams from P_2 thus indicating that there is a slight undulation around the north-south axis in the vertical north-south plane, a conclusion that is supported by the comparison of the P and R diagrams.

The two collective diagrams made from 330 and from 305 lamellae show, within the occupied 100° arc, and within the measurable field in P, a definite concentration of poles into one dominant maximum A close to the east-west pole. Another distinct submaximum B is about 45° above A. On the periphery of the diagram there are two submaxima C and D. The fact that these maxima persist in approximately the same place and approximately the same shape when the count is stepped up from 150 to 500 poles indicates that they represent significant rather than accidental concentrations of lamellae. This conclusion is further confirmed by the fact that even in as low a count as 90 poles, the general position of A and B is outlined by an area of relatively high concentration. The detailed pattern of A, and the submaxima C and D appear when more poles are counted.

The upper left hand quadrant in all the diagrams of P, i.e., the upper north-west quadrant indicates a northward pull of the arcuate area of high concentration. This is shown by the recurrence of a sub-maximum H of varying intensity inside the peripheral zone of the diagram. However, the fact that H does lie *inside* the periphery and within the transition zone of incomplete counting makes it impossible to draw correct analytical conclusions from the position of H until the diagram is filled. The only conclusion that can be safely drawn from the unfilled diagram of P is that in the field of the thin section and of the thin section block, the fabric is essentially homogeneous and that there are two significant maxima and one or two submaxima restricted to a peripheral arc passing through the east-west pole and extended upward and downward on both sides of this pole.

R_1 and R_2 diagrams (unfilled collective). Elemental and

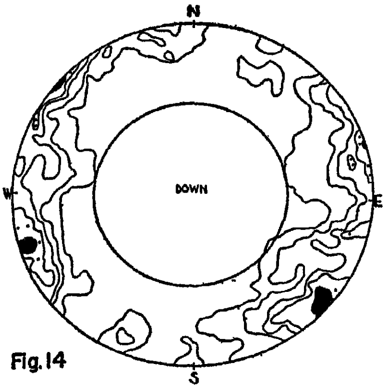

Fig. 14

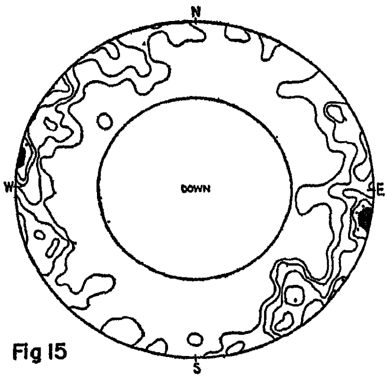

Fig 15

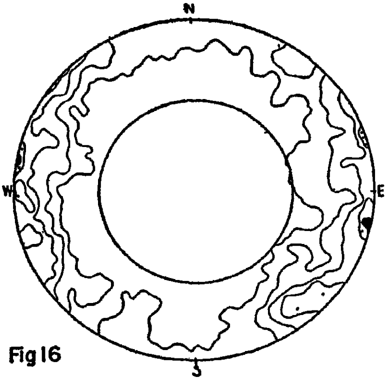

Fig 16

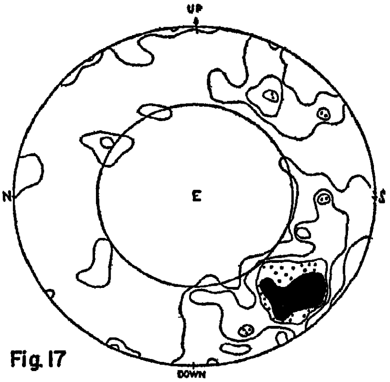

Fig. 17

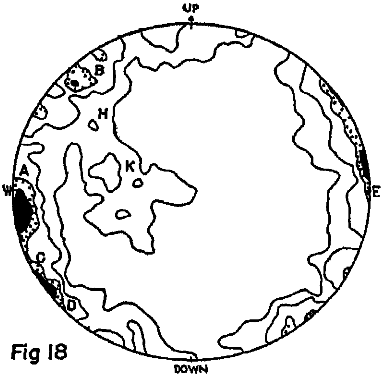

Fig 18

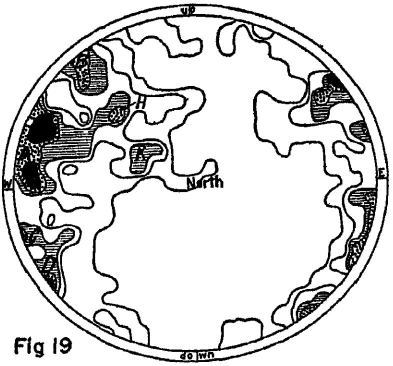

Fig 19

Figure 14. 292 poles of calcite lamellae in R_1 section. Contours (6-5)-4-3-2-1-0%.

Figure 15. Collective diagram of 203 poles of calcite lamellae in R_2 section. Contours (7-5)-4-3-2-1-0%.

Figure 16. 494 poles of calcite lamellae in R_1 and R_2 sections. Contours (6-5)-4-3-2-1-0%.

Figure 17. 190 poles of calcite lamellae in Q section Contours (6-5)-4-3-2-1%.

Figure 18 410 poles of calcite lamellae in P_2, 325 poles measured in P_2, 85 poles rotated from R_2 to fill P_2. Contours (6-5)-4-3-2-0%.

Figure 19. 385 poles of calcite lamellae in P_1 filled by rotation of 80 poles from Q. Contours (5-3.5)-3-2.5-2-1-0%.

collective diagrams from the R_1 section, cut from the horizontal plane and mounted so that looking down in the thin section is looking down in Nature, show, near the emergence of the east-west axis, the same concentration of poles that characterized the diagrams from P_1. There is also a northward and southward extension of the peripheral zone of high concentration forming an arc of 80°-90° on both sides of the east-west axis, that shows plainly in the diagram of 292 calcite lamellae from R_1 (fig. 14). A minimum near the north and south poles in the R diagram could never appear in the diagram measured from P because the area around the north and south pole is unmeasurable in P. The same minimum shows clearly in the diagram measured by Dr. Bell from Q, (fig. 17). The general pattern of the calcite lamellae, deduced from the two diagrams from P and from R, is a concentration common to both diagrams, covering a polar cap of the projection sphere of about 30° radius. The A maximum in P reappears in R. The B and D of P fall outside of the measurable area in R and therefore do not appear in the R diagram.

The diagram from the thin section R_2 was contoured first in two elemental diagrams from two traverses in different parts of the thin section. The field of the thin section proved to be reasonably homogeneous. Therefore, all the poles measured in R_2 were contoured in a collective diagram of 203 lamellae (fig. 15). This collective diagram of 203 lamellae in R_2 coincides so well with the collective diagram of 292 lamellae in R_1 that a collective diagram of 494 lamellae from R_1 and R_2 (fig. 16) represents the field of the whole thin section block in the horizontal plane. The same pull from the periphery toward the vertical axis that showed in the quadrant delimited by the north, west and vertical axis in the diagram of P_1 at H reappears in the diagram of R_1 in the corresponding quadrant delimited by the south, east and vertical axis. But, as in the diagram from P_1, so in the R_1 diagram it is impossible to evaluate the significance of this area of high concentration because the true relative concentration can only be found from analysis of the Q_1 section.

Q_1 diagrams (unfilled collective). Two diagrams of 100 and 190 poles of calcite lamellae each made from the Q_1 section looking east (fig. 1,a) coincide so closely that the field of the thin section can be considered homogeneous. Figure 17,

which is a diagram of 190 lamellae shows a strong concentration in the lower southeast quadrant. This maximum seen in diagrams from Q is clearly indicated at H in the diagrams from P and from R although it can only appear in its entirety in the diagram from Q. There is also a submaximum beginning to appear in the upper southeast quadrant in the 190 pole diagram. But more than one half of the field of the projection is unoccupied in Q, which is what is to be expected on account of the strong concentration at A around the emergence of the east-west axis in the P and R diagrams, a concentration that could never appear in Q diagrams. Therefore, the measured maximum is over-emphasized in Q because of the absence of the maxima that are so prominent in the diagrams from P and R. For this reason the diagram measured from Q is of comparatively little value in interpreting the preferred orientation until Q is filled, both from P and from R.

FILLED DIAGRAMS

The three diagrams from the P, R and Q sections were filled respectively from sections parallel to the two faces at right angles to P, R, and Q, in order to check more closely the slight inhomogeneity in the field of the thin section block that is shown by the unfilled diagrams. Lamellae diagrams from P filled from R and Q are shown in figure 18 and figure 19. The correspondence between the P diagrams filled from R and from Q is good although the lamellae maximum at the west pole of the diagram is above the west pole in P_1 and below the west pole in P_2, a fact that is explicable by the oscillation of the axes maximum on either side of the east-west pole. A complete collective diagram was then made from 756 poles from P_1 and P_2, filled from R_1 and R_2 (fig. 20) for comparison with 656 poles from R_1 and R_2 filled from P_1 and P_2 (fig. 21). Figure 22, which is a lamellae diagram measured in Q and filled by poles rotated both from P and from R shows that the dominance of the east-west maxima has relegated the maximum of the unfilled Q diagram to a submaximum.

Then 385 lamellae poles from P filled from Q (fig. 19) were rotated to the position of R (fig. 23) for comparison with figure 21 (656 poles from R_1 and R_2 filled from P_1 and P_2). The correspondence between these lamellae diagrams in the areas common to both P and R is good. The correspondence between the axis diagram measured in P and rotated to R (fig. 8) and the diagram measured in R (fig. 7) is also good and the essential homogeneity of the fabric is established.

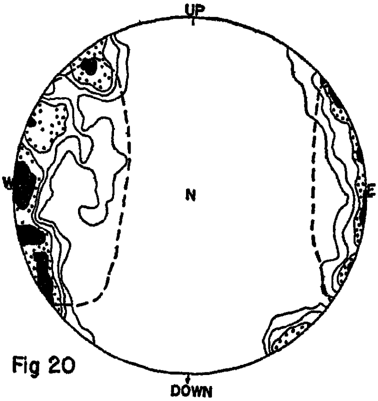

Fig 20

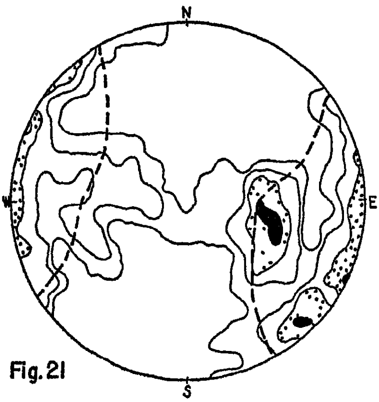

Fig. 21

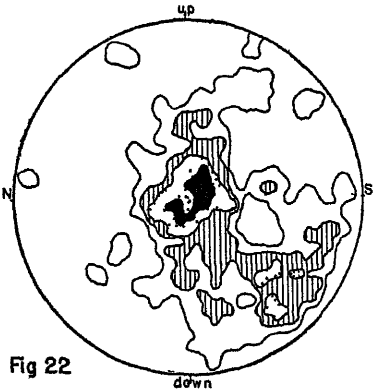

Fig 22

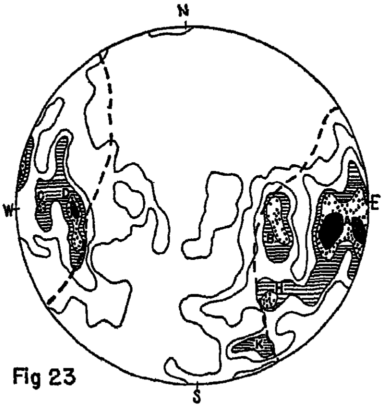

Fig 23

Figure 20. 756 poles of calcite lamellae: 305 poles from P_1, 315 poles from P_2, filled with 52 poles from R_1 and 84 poles from R_2 Contours (5-4)-3-2-1-0%

Figure 21 656 poles of calcite lamellae from R_1 and R_2 filled from P_1 and P_2. Contours (5-4)-3-2-1-0%.

Figure 22. 838 poles of calcite lamellae in Q section: 172 measured in Q, 87 rotated from P, 78 rotated from R. Contours (6-4)-3-2-1-0%

Figure 23 385 poles of calcite lamellae in P_1 filled from Q and rotated to the position of R. Contours (5-3.5)-3-2.5-2-1-0%

INTERPRETATION OF THE LAMELLAE DIAGRAMS

Both P and R diagrams indicate a scattering of maxima within the respective fields of *high concentration* (Überbesetzung), i.e. a certain multiplicity of maxima. Scattering of maxima in lamellae diagrams may mean one of two things:

(1) an external rotation of the fabric has involved a flexure, which entails a relative displacement of the mass as a whole. If calcite lamellae or cleavage planes of mica are arranged parellel to the *s* plane in a flexural slip fold, the poles within the field of that flexure will occupy positions that are definitely oriented with respect to the orientation of the whole mass but which are different in sections cut from different positions within the flexure.[22] The resultant fabric will be inhomogeneous and as the fabric of the Yule marble is essentially homogeneous the various maxima in the lamellae diagrams do not indicate flexural slip in the fabric. (2) An internal rotation has resulted in the formation of two or more interesting *s* planes along which the fabric elements are alined. The resultant fabric will be homogeneous.

Calcite lamellae can result from the polysynthetic development of mechanical twinning along planes of displacement that are parallel to $(01\bar{1}2)$. This method of twin-gliding can be produced in the laboratory by the so-called Baumhauer experiment. Bell[23] has also induced translation-gliding in calcite by applying a force in such a way as to inhibit movement in the sense of the twin-glide direction in $(01\bar{1}2)$. Therefore in his experiment the lattice could not deform by twin-gliding but the calcite did deform by translation-gliding in a homogeneous strain with the development of lamellae parallel to $(01\bar{1}2)$ in a manner analogous to the production of slip bands in metal working. See Plate IV.

Sharp lines representing traces of surfaces of discontinuity in the grains can also be the traces of rhombohedral planes parallel $(0\bar{1}11)$. Therefore, it is important to know the crystallographic relation of given maxima to the optic axes maxima. This relation can be determined by measuring the optic axis in each grain together with the trace of all the lamellae in this grain. Considerable drawback to exact measurement of both elements in the same grain is the fact that relatively thick sections give the most satisfactory results in placing the trace of lamellae in the vertical position, whereas relatively thick sections are eminently unsatisfactory for the accurate placing of optic axes in a measurable position in a mineral of such high birefringence as calcite. Unfortunately, in the first work on the undeformed fabric the axes and lamellae were measured separately, so they cannot be directly related, grain by grain. However, in later work grain measurements were recorded grain by grain. Then the locus of *e* planes {0112} that could belong to the axes occupying the maximum in the axis diagram was plotted. The lines marking the outer limit of possible *e* planes related to the axes maximum are shown by the heavy dashed line in figures 20, 21 and 23. They coincide so well with the outer limits of the lamellae maxima that probably most of the measured lamellae represent *e* planes. This is confirmed by the evidence of those grains in which axes and lamellae were measured grain by grain.

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[22] Knopf, E. B., and Ingerson, E Structural Petrology. Geol. Soc. America. Mem. 6, 77-79, 1937.

[23] Bell, James F. Personal Communication.

GENERAL CONCLUSIONS IN REGARD TO THE FABRIC OF THE YULE MARBLE BEFORE DEFORMATION

A megascopic *s* plane, striking N5°E and dipping almost 90°, is shown by the "grain", i.e. by the dimensional orientation of the calcite grains. Petrofabric analysis indicates that this plane contains a strong concentration of lamellae into maxima, occupying in general the position that would be occupied by *e* planes belonging to the optic axes shown by the maxima in the optic axes diagrams.

It has long been known that *e* planes in calcite can act as operative[24] gliding planes in the calcite lattice, and the fact that these crystal planes lie in or near to the megascopic *s* plane of the fabric suggests that intragranular gliding movement along these *e* planes may have brought a large proportion of the calcite grains into their present preferred orientation.

In terms of petrofabric analysis the parallel surfaces along which the orienting of slip proceeds are defined by the axes *a* and *b*. Therefore the northward trending vertical plane occupied by the twin or translation lamellae would be the *ab* plane. By definition the axis *a* represents the direction of the movement and the axis *b* represents axis of slip, which is the normal to the movement direction. In the megascopic

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[24] Operative is used here to denote the plane on which slip has occurred during the orienting process by which the fabric of the Yule marble was formed. Cf Gough, H. J.: Crystalline structures in relation to the failure of metals, especially by fatigue. Am. Soc. for Testing of Materials, 33, pt. 2, 18, 1933.

s plane (*ab*) there is nothing to determine the relative positions of *a* and *b*. The normal to the slip plane is the fabric axis *c*, which corresponds to the east-west direction and to the main lamellae maximum in the diagrams. The main slip plane of the fabric, (*ab*), established by the lamellae maxima determined in fabric analysis, coincides with the megascopic "grain" of the fabric.

The fact that the ratio of the longer to the shorter grain axes in sections parallel to R is 3 to 1 as contrasted with 1.8 *to 1 in the section parallel to P suggests that the maximum* extension of material is in the north and south direction, which would be the fabric axis *a*. However confirmation of the position of *a* would require exact determination of the crystal planes occupying each lamellae maximum. This has not been done in this preliminary work and the present work indicates only that the directions joining the lamellae poles and the optic axes are in or close to the horizontal plane in nature, a conclusion that is supported by the orthorhombic symmetry in the test cylinder after deformation, which is described in Part II.

The position of the center of gravity in the dominant axes maximum on *one* side rather than on both sides of the fabric axis *c* establishes the monoclinic symmetry of the undeformed fabric.

In order to evaluate the changes in fabric that were produced by compression in the laboratory the existence of a distinct *s* plane striking nearly north and standing approximately vertical, together with a statistical preference of the optic axes for a position that is inclined at 64° to the *s* plane, is the important feature of the undeformed fabric. Recrystallization — consequent upon thermal metamorphism could expectably emphasize this *s* plane. Whether this recrystallization is a paracrystalline growth accompanying a movement of intragranular slip or whether it is a mimetic recrystallization recording a pre-existing anisotropy could only be decided after more detailed study of the fabric under the microscope.

For the present study of the effect of compression, the requirement of a comparatively simple internal grain fabric *that can be definitely related to the direction of applied force* in deformation is fulfilled by the fabric of the Yule marble.

(*To Be Continued*)

TWO ENDOCRANIAL CASTS OF CETACEANS FROM THE OLIGOCENE OF NEW ZEALAND

B. J. MARPLES

ABSTRACT. Two natural endocranial casts of Cetaceans from the Oligocene of New Zealand are described. No skeletal remains are associated with either of them. The larger closely resembles published figures of endocranial casts of Archaeocetes except in that its cerebellum is small and normal instead of being of immense size The cerebral hemispheres are small, especially anteriorly, and apparently had an olfactory region. The smaller specimen more closely approaches the condition seen in Recent specimens. The cerebral hemispheres have no olfactory region and are rounded anteriorly, but they are far less swollen The two specimens afford intermediate stages between the brain of the primitive mammal and that of the Recent Odontocete. It is suggested that the immense and unique cerebellum described previously in the brain of Archaeocetes might conceivably be a misinterpretation of a large vascular plexus.

IN the collection of the Otago Museum, Dunedin, New Zealand, are two natural endocranial casts belonging to Cetaceans from the Early Tertiary. The larger one (No. C.48.70) has no data attached to it, while the smaller (No. C.34.7) was collected at Milburn, Otago. There seems no doubt but that both are of Oligocene age. They were mentioned in a paper dealing with *Prosqualodon hamiltoni* by Benham, who pointed out that the larger one is roughly similar in size to the cranial cavity of the skull of this Squalodont, though the skull is not well enough preserved for any of the details to be compared. The only literature on the endocranial casts of cetaceans available has been the papers of Elliot Smith and of Dart on the brains of Archaeocetes. Dart also figures and discusses an endocranial cast of *Prosqualodon davidi* from the Miocene of Tasmania. The present specimens are so unlike those in the literature that it seems desirable that they should be described, and I am indebted to the Director of the Otago Museum for permission to do so.

The larger specimen (Fig. 1), as preserved, is 145 mm. long, 148 mm. wide and 89 mm. high, and it has a volume of approximately 670 cc. It is broken off transversely at the anterior end and obliquely in the region of the medulla, while a portion is missing on the right side. The rest of the surface is well preserved and the impressions of many bloodvessels are clearly visible. The broken surface at the anterior end forms a roughly equilateral triangle with sides about 50 mm. long. The dorsal angle runs back for some 35 mm. as a sharp ridge, the cerebral hemispheres swelling gently on either side.

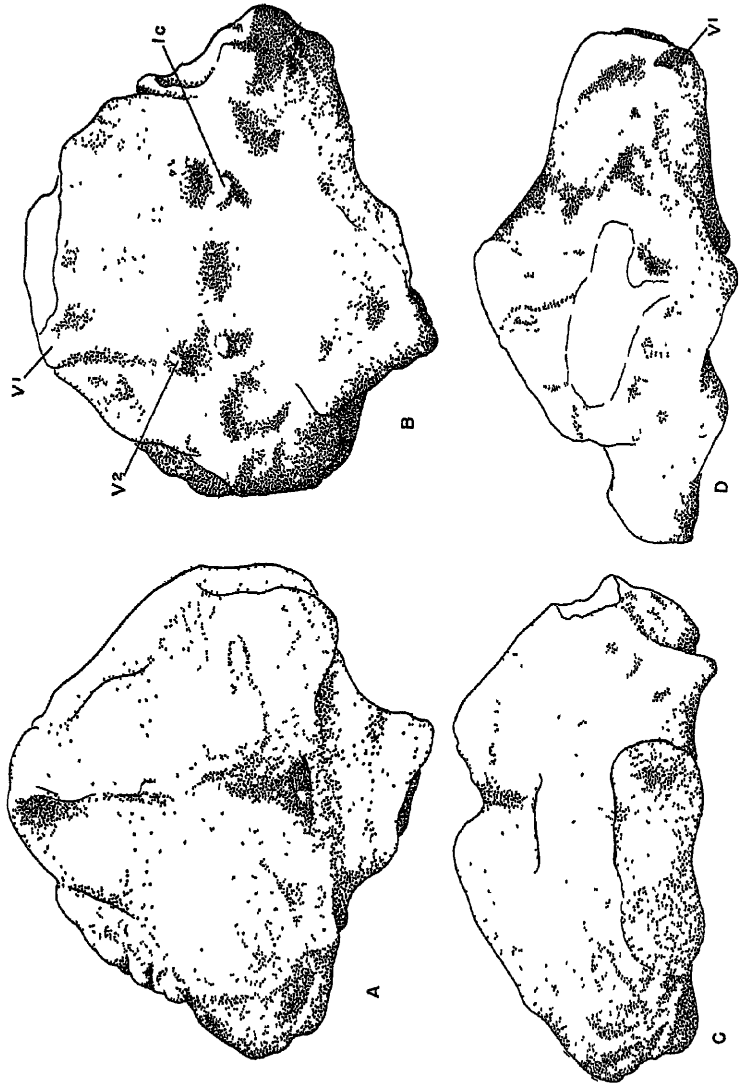

Fig. 1. Natural endocranial cast of Oligocene Cetacean from New Zealand. Larger specimen, No. C 48.70 Actual length 145 mm. A Dorsal side. B Ventral side. C Posterior side D. Right side I.C Internal carotid V 1. Maxillary branch of trigeminal nerve. V.2 Mandibular branch of trigeminal nerve

Indications of the sagittal sinus are visible in the middle line posterior to the ridge, and it extends into a deep V-shaped groove between the posterior regions of the cerebral hemispheres, which rise on each side into rather sharp conical projections. Still more posteriorly the median groove widens out into a triangular depression bounded behind by a sharp transverse ridge. This depression presumably indicates the position of the tentorium cerebelli.

The cerebral hemispheres are low and narrow anteriorly, but expand rapidly both sideways and upwards towards their posterior ends. A shallow lateral depression separates a slight anterior swelling from the high conical region behind. The surface of the posterior half, as far forward as the highest peak, is covered with a network of bloodvessels, some of which are indicated in the figures. The best marked is a vessel which takes a curving course outwards from the sagittal sinus, passing posterior to the peak of the cerebral hemisphere. A larger, but less clearly defined vessel, runs from the region of the ridge behind the tentorial depression anteriorly round the lateral corner of the cerebral hemisphere. The details on the sides of the cast are not clear. Lying alongside the anterior half of the cerebral hemisphere is an irregular mass which seems to be composed of bloodvessels and appears to be connected with the posterior cerebral vessel and below with a large vessel which emerges between the cerebral hemisphere and cerebellum. This irregular mass is separated medially from the cerebral hemisphere by a groove, and is divided into three lobes by two oblique dorso-ventral grooves. Small fragments of bone still adhere to the cast in these grooves. The anterior end of the lateral irregular mass is broken off and lies adjacent to the corner of the anterior triangle as a small triangle about 10 x 15 mm.

The posterior region of the cast is broken off obliquely and only extends some 45 mm. behind the transverse ridge. From the ridge the dorsal surface sweeps smoothly down with only a slight transverse swelling and a few fine longitudinal bloodvessels. The cerebellar region extends obliquely downwards and forwards under the overhanging posterior edge of the cerebral hemisphere.

The ventral surface of the cast is broad and flat, its most conspicuous feature being a low transverse swelling slightly anterior to the middle. Posterior to the ends of this are small elevations apparently showing the positions of the internal carotid foramina, and lateral to these the large bloodvessels come from the groove between the cerebellum and cerebral hemispheres and turn outwards towards the lateral irregular mass of bloodvessels. Smaller vessels appear from the same groove and also run laterally. Lateral to the ends of the transverse swelling are elevations probably representing the foramina for the mandibular branch of the trigeminal nerve, while the maxillary branch and the optic nerve appear to run forward at the lower corner of the anterior triangle.

Two artificial endocranial casts of Recent Odontocetes were available for comparison, one of *Lagenorhynchus obscurus* and the other of *Cephalorhynchus hectori* (fig. 2, E & F). A number of striking differences are at once apparent. In the fossil the main axis is straight while in the Recent specimens the anterior half is bent upwards at an angle of roughly 40°. The flatness of the ventral surface of the fossil is the more noticeable in that the lateral lobes of the cerebellum hardly extend below the ventral surface, while in the Recent casts they are much larger and more rounded and extend 15-20 mm. below it. The whole of the Recent cerebral hemisphere is more rounded and the anterior end is without the prismatic appearance of the fossil, while there is no sign of the lateral irregular mass of bloodvessels. Apart from these differences the fossil and Recent casts are easily comparable and the cerebral bloodvessels have the same course. The vessel emerging between the base of the cerebral hemisphere and the ventral anterior corner of the cerebellum runs round the back of the cerebral hemisphere on to the cerebellum near the tentorium and thence round to the ventral side of the medulla, while the middle cerebral vessel runs round over the apex of the cerebral hemisphere to the sagittal sinus. On the ventral surface, the internal carotid foramen and that for the mandibular branch of the trigeminal nerve are similarly situated.

The differences between the size and shape of the cerebral hemispheres and cerebellum in the fossil and the Recent specimens are such as one would expect in comparing the brains of Oligocene and Recent members of the same group of mammals. The difference at the anterior end seems to be due to the loss of the olfactory sense in the Recent Odontocete and perhaps also to the telescoping of the skull.

On comparing the fossil cast with the figures given by Elliot Smith and Dart of the endocranial casts of *Archaeocetes*, a very strong general resemblance is at once apparent. The cerebral hemispheres of the Archaeocete are low and rounded anteriorly, and there is a dorsal ridge leading to a prismatic anterior end, formed by the olfactory peduncles and anterior nerves. On each side is an irregular mass, interpreted by Dart as a huge Gasserian ganglion covered by middle cerebral vessels. The posterior part of the Archaeocete cast rises to a peak, superficially much as does the present specimen.

The peculiarity of the Archaeocete brain lies in the cerebellum. This is an irregular transverse mass rising high above the cerebral hemispheres and extending down the sides in what are referred to by Dart as the parafloccular lobes. To quote Elliot Smith, "perhaps the most striking feature of the brain of *Zeuglodon* is the extreme disproportion between the size of the enormous cerebellum and the diminutive cerebrum. In this respect the fossil brain presents a most marked contrast to that of all Recent mammals, and especially to that of the Cetacea." Much discussion has centered on the size of the cerebellum, and its relation to the systematic position of the Archaeocetes and their mode of life.

As pointed out above, if this information had not been available no difficulty would have been experienced in comparing the specimen with the Recent casts, its cerebellum being quite similar though relatively smaller (fig. 2). In view of the published information about the Archaeocete cerebellum and the close similarity in shape between the Archaeocete cast and the present one, a close reexamination was made. There seems no doubt but that the whole of the anterior portion of the cast represents the cerebral hemispheres. A smooth continuous surface is visible with no trace of a transverse tentorium in front of the highest part, while there is one posteriorly, exactly as in Recent specimens. The elevated portions are separated by a deep sagittal groove as one would expect if they are cerebral hemispheres but not if they are cerebellum, and this is also visible in some of Dart's figures. The relations of the bloodvessels are also exactly as in the Recent specimens.

Unfortunately nothing is known of the origin of this specimen and Archaeocetes, Squalodonts and Cetotheres are all present in the Early Tertiary of New Zealand. The skull of *Prosqualodon hectori* is not sufficiently well preserved to show what was the shape of its brain, and the endocranial cast of

Prosqualodon davidi figured by Dart is quite unlike the present specimen. The balance of evidence, except for the condition of the cerebellum, seems to point to the present cast belonging to an Archaeocete. If it could be shown that the cerebellum of the Archaeocete was in reality a small normal one, the history of the Cetacean brain would be much simplified and the present specimens would fit readily into it. It is of course impossible to determine anything definite without examination of the specimens, but several points in the published figures and descriptions suggest that an alternative explanation is conceivable. The peculiarity of the Archaeocete structure is exemplified by Elliot Smith's statement, "behind the part which I have just described as the cerebrum there is a large irregular mass of a very peculiar shape, not exactly comparable with the condition occurring in any other brain known to me." Close examination, however, convinced him that it formed part of the cerebellum. Examination of Dart's figures shows

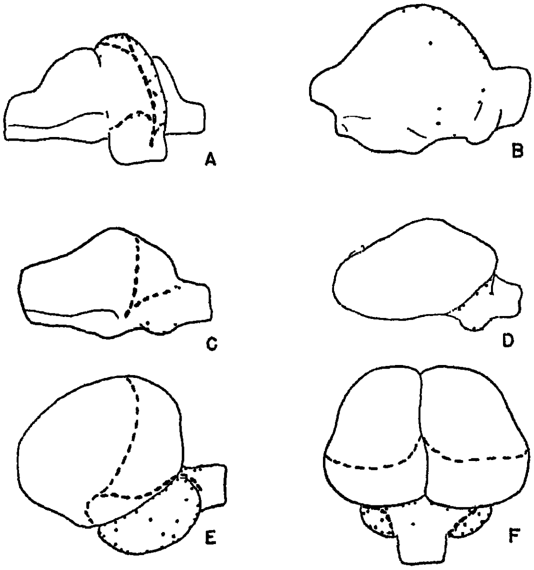

Fig. 2. Diagrams of endocranial casts of Cetaceans reduced to approximately the same size. Cerebellar region dotted, bloodvessels in broken lines. A. Left side of *Zeuglodon osiris* (bloodvessel from *Prozeuglodon atrox*), after Dart. Middle Eocene. B Left side of *Prosqualodon davidi*, after Dart. Miocene. C. Left side of larger specimen Oligocene. D Left side of smaller specimen Oligocene. E. Left side of *Cephalorhynchus hectori*. Recent. F. Dorsal side of *Cephalorhynchus hectori*. Recent.

that this region has a very irregular surface, due to the presence of either numerous small sulci or of bloodvessels. In some specimens there is a deep sagittal groove. It seems possible that this mass represents, not an immense anterior development of the cerebellum, but one of the retia mirabilia which are known to occur in Recent Cetaceans. Ommanney, describing these structures in the cervical region of *Balaenoptera* states that "up to the foramen magnum the neural canal is filled with a plexus, which is entirely venous, embracing the spinal cord and entering the skull through the foramen magnum to form a vascular mass against the hinder surface of the brain." In support of this suggestion is the fact that the corresponding region of the present specimen is covered with the impressions of bloodvessels, though not so thickly as to obscure the outline of the cerebral hemispheres. In Dart's figure of *Prozeuglodon atrox* the transverse sulcus anterior to this mass is very slight, while a bloodvessel runs over its apex very much as one does over the cerebral hemisphere of the Recent cast and of the present specimen. Further support to the vascular theory is the statement by Kellogg that the vertebrarterial foramina in the cervical vertebrae of *Basilosaurus cetoides* are very large, and that the main arterial supply to the brain of Recent Odontocetes is by this route.

The smaller specimen (fig. 3) is 88 mm. long, 89 mm. wide and 50 mm. high, and is asymmetrical, being strongly skewed over to the right. Its volume is approximately 200 cc. The cerebral hemispheres are low anteriorly, rising to a rounded elevation behind, the highest part being much less acutely pointed than in the larger specimen. The surface is smooth and without bloodvessels except for a few small ones entering the sagittal sinus. The anterior ends of the cerebral hemispheres are rounded and between them is a striking hexagonal projection whose concave anterior face contains traces of bone. No comparable depression was observed in several Recent porpoise skulls, and the shape of this part of the cast is quite unlike that of the larger one. Dart figures a conical anterior projection on the cast of *Prosqualodon davidi*, which he refers to as the olfactory peduncle. On the left side is a large mass somewhat ridged on its surface in a way suggestive of bloodvessels. There are traces of bone in the groove between this mass and the cerebral hemisphere. No transverse tentorial depression is visible posterior to the cerebral hemispheres and the details of the posterior region are obscure.

The ventral surface is well preserved as a smooth undulating surface, all approximately in one plane. In the centre are several elevations and depressions, the largest apparently

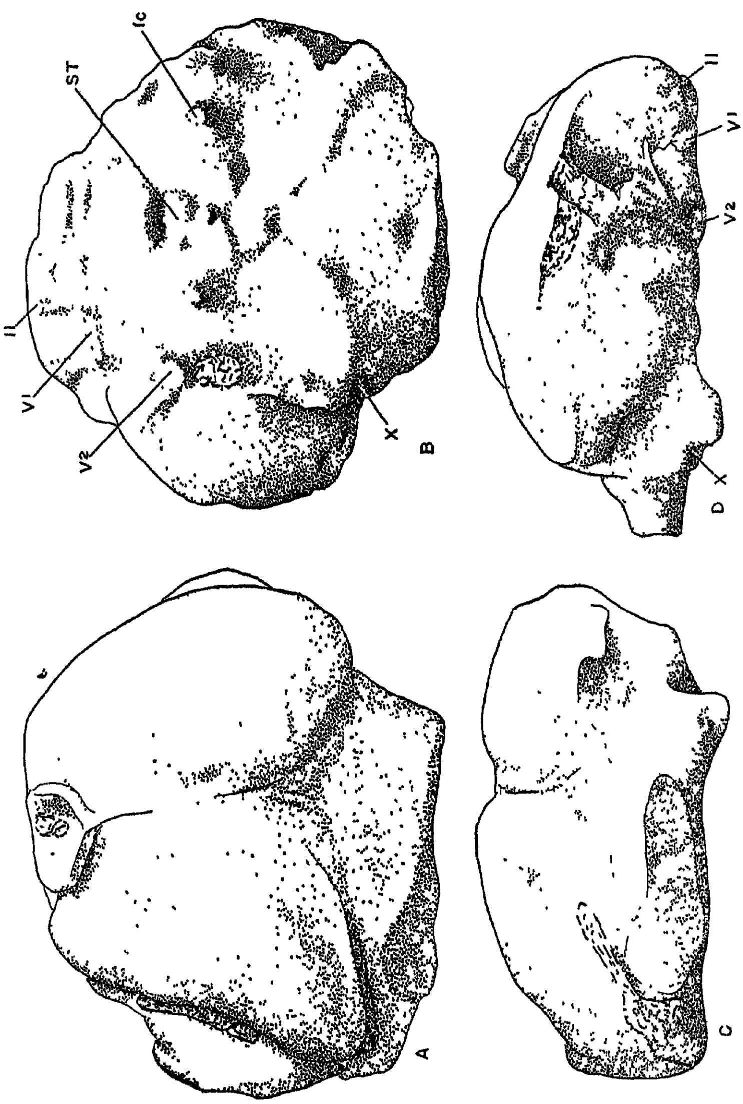

Fig 8. Natural endocranial cast of Oligocene Cetacean from New Zealand. Smaller specimen, No. C.84 7. Actual length 88 mm A Dorsal side. B Ventral side C. Posterior side. D. Right Side. S.T. Stella turcica I.C. Internal carotid. II. Optic nerve. V 1. Maxillary branch of trigeminal nerve V 2. Mandibular branch of trigeminal nerve. X Vagus nerve

representing the sella turcica, and laterally there are elevations indicating the carotid foramina. Postero-laterally there is the elevation below the cerebellum and on its posterior surface is the impression of a foramen, possibly that for the vagus nerve. Anterior to this elevation is a deep depression containing fragments of bone on the right side of the cast, and anteriorly again are two elevations apparently representing the foramina for the mandibular and maxillary branches of the trigeminal nerve. At the anterior end on the right side can be detected what is probably the foramen for the optic nerve. The arrangement of these features is not unlike that seen on Recent endocranial casts, though in these they are more crowded together.

The smaller cast seems to represent a brain more closely resembling that of a Recent Odontocete than does the larger one. It differs from the Recent endocranial cast in being flat on the ventral surface, in having a relatively smaller cerebellum and very much less expanded cerebral hemispheres. It differs from the larger cast principally in the more rounded cerebral hemispheres especially at the anterior end where the cast is not drawn out into the prismatic anterior region. A remarkable feature is its very small size, 200 cc., and it seems likely that it belonged to a young individual. All the known New Zealand Odontocetes of camparable geological age are considerably larger, and according to Dart the volume of the endocranial cast of *Prosqualodon davidi* from the Miocene of Tasmania, is approximately 750 cc. Dart comments on the fact that this volume is slightly less than that of the endocranial cast of the Eocene *Zeuglodon intermedius*. He maintains that this is sufficient, in accordance with the "law of increasing brain weight" put forward by Marsh, to show that the Zeuglodont cannot have been ancestral to the Squalodont, but that the ancestor must have had a brain capacity very much less. Edinger, however, as a result of a study of the phylogeny of the horse brain, criticizes Marsh's law.

The larger specimen is interesting in the light of Edinger's findings in connection with the evolution of the horse brain. She showed that in *Eohippus*, where the skeletal characteristics already clearly foreshadowed the line of descent leading to the horse, the brain was completely unspecialized and comparable with only the most primitive of living mammals, such as the opossum *Didelphys*. As she says, "the outstanding characters of the skeleton place *Eohippus* with the Equidae; the osteologist can see the future horse in this Lower Eocene form. The brain connects *Eohippus* only with the past. The ancestral brain, an unspecialized mammalian brain, was held over in the Equid body. No features of this brain signal the future." The cerebral hemisphere of the larger cast almost resemble those of *Eohippus*, though more expanded posteriorly so that the mid brain and cerebellum are less conspicuous. Assuming that the larger cast belongs to an Archaeocete, it is tempting to speculate as to whether the specialization of the bodies of these whales for aquatic life had taken place while the skull and especially the brain remained in a primitive condition. In the Odontocete, telescoping of the skull, loss of olfactory sense and enlargement of the cerebral hemispheres seems to have already taken place by Oligocene times, to judge from the smaller specimen, while in Recent times the cerebral hemispheres and cerebellum are greatly enlarged and exhibit the greatest degree of convolution of any brain. It is clearly recognized, however, that this interpretation of the present specimens does not fit in with those of Elliot Smith and Dart for the Archaeocete and Squalodont endocranial casts at their disposal, and further material must be awaited.

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DEPARTMENT OF ZOOLOGY
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STRATIGRAPHY AND TRILOBITE FAUNAL ZONES OF THE GARDEN CITY FORMATION, NORTHEASTERN UTAH

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ABSTRACT. The Garden City formation is widely distributed in northeastern Utah and the adjacent portions of southeastern Idaho, ranging in thickness from about 1,200 feet on the east to 1,800 feet on the northwest of the Logan quadrangle. As a lithologic unit it is easily distinguished from the dolomite of the underlying St. Charles formation and from the shales, siltstones, and quartzites of the overlying Swan Peak formation.

The Garden City beds can be divided into two members, the lower, comprising approximately two-thirds of the formation, being composed of numerous alternations of interbedded and interlensed intraformational conglomerates. and crystalline, aphanitic, and muddy limestones. The upper member is characterized by its high content of black chert nodules, stringers, and "interbeds" occurring for the most part in irregularly laminated, aphanitic limestone and dolomitic limestone

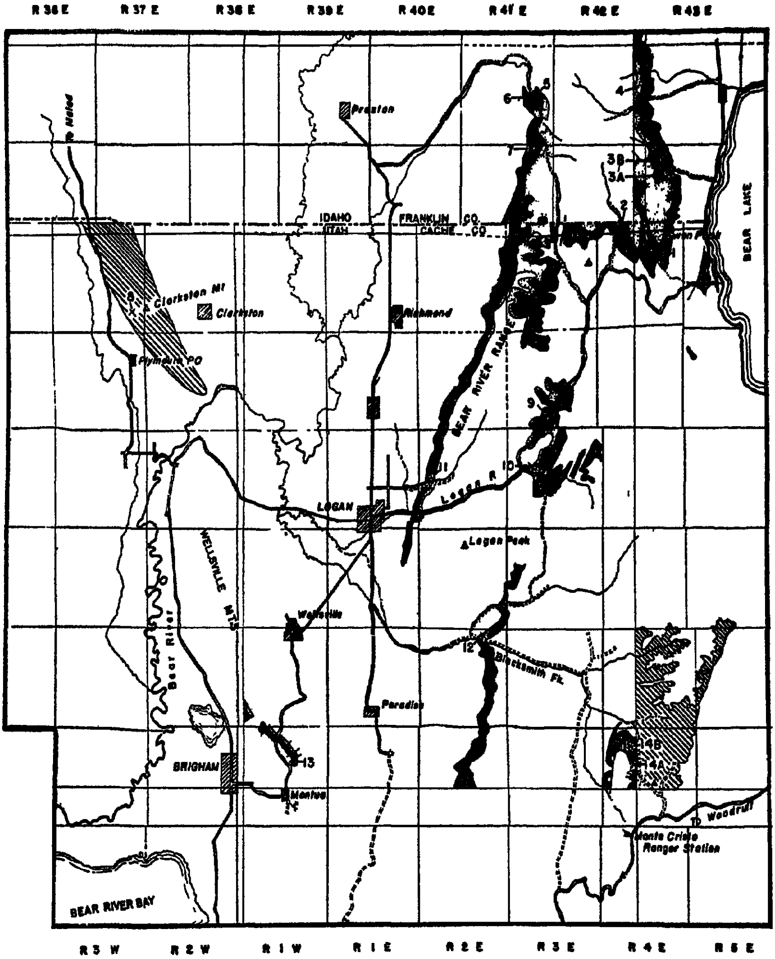
Faunal zones near the base and top of the formation indicate that its lower and upper boundaries vary little, if at all, in age within the area studied. A sequence of 12 faunal zones, distributed throughout the formation, indicate it to be mostly Canadian in age and contemporaneous with at least a part of the Pogonip limestone of Nevada The upper 30 to 50 feet are faunally allied with the overlying Swan Peak formation and are almost certainly of Chazyan (Joins) age. On the basis of trilobites the lower 60-150 feet (including faunal zone "B") are correlated with the Tribes Hill and Stonehenge formations of the East.

The prolific trilobite fauna of the Garden City beds includes more than 80 species, detailed descriptions of which will be presented in a future publication.

The Swan Peak formation, now known to be composed of a more varied lithology than the original definition suggests, is over 550 feet thick in the northwest corner of the area studied, wedging-out completely to the southeast. Although the nature of its contact with the Garden City strata is apparently gradational to some degree, evidence for the direction of transgression of the younger detrital sediments is not clear in this region; such evidence must be sought in the western Utah and Nevada sections of the Pogonip formation and Eureka quartzite.

INTRODUCTION

IN 1913 the name "Garden City limestone" was given by Richardson to strata of Early Ordovician age in northeastern Utah and southeastern Idaho. These beds, which bear ample evidence for shallow water deposition, vary in thickness from 1,200 feet to 1,800 feet. As a result of the present study, to date limited to 14 strategically located sections within the area bounded by parallels 41°30′ and 42°10′ and meridians 111°20′ and 112°20′ (fig. 1), it has been possible to divide the formation into two lithologic members. A special search



AREAL DISTRIBUTION OF THE GARDEN CITY
AND SWAN PEAK FORMATIONS
in Logan and Randolph Quadrangles, Utah,
and southern Montpelier and Preston Quadrangles, Idaho.
Scale 0 1 2 3 4 5 6 7 8 9 10 MILES
BOISE AND SALT LAKE MERIDIANS

■ GARDEN CITY FM ☐ SWAN PEAK FM ▨ UNDIFFERENTIATED LOWER PALEOZOIC

COMPILED FROM
BUTLER ET AL 1920
MANSFIELD 1927
RICHARDSON, 1941
WILLIAMS 1948

Figure 1. Index map of the area studied, indicating the positions of stratigraphic sections and collecting localities.

for silicified fossils, suitable for etching, has resulted in the discovery of more than 80 species of trilobites; these with other elements of the fauna are the basis for the designation of 12 faunal zones within the formation. Because of the nature of its contact with the Garden City beds, some attention was given to the overlying Swan Peak formation.

As originally defined by Richardson (1913, p. 408-409) with its type section in Garden City Canyon, Sections 3, 4, and 10, Township 14 North, Range 4 East, the Garden City formation was stated to consist of a "succession of thick and thin bedded gray limestone approximately 1,000 feet thick" and to be characterized by the "presence throughout the formation of a conglomerate or breccia consisting of elongate bits of limestone up to 2 or 3 inches in length, irregularly imbedded in a matrix of similar composition." Richardson further noted that the formation apparently overlay the St. Charles formation disconformably and was overlain by the Swan Peak "quartzite" conformably (1913, p. 408). It was considered to be of Beekmantown age.

Mansfield recognized and mapped the formation in the Fort Hall Indian Reservation (1920, p. 32 and Plate III), in the Portneuf Quadrangle (1929, p. 19), and in the Montpelier Quadrangle (1927, p. 57) in Idaho. In his report on the last of these he increased Richardson's thickness to 1,250 feet. In 1934 Kirk (p. 456) suggested the contemporaneity of the El Paso limestone with the Garden City beds.

Because of a series of confusions, stemming from erroneous interpretations of the relation between the St. Charles and Garden City formations (Williams, 1948, p. 1135) Richardson (1941, pp. 13-14) incorrectly increased the thickness of the formation in the Randolph Quadrangle to 1,900 feet. This error is rectified by Williams in his recent account of the Paleozoic rocks of the Logan Quadrangle (1948, pp. 1135-1137); in his report Williams notes the presence of "thick beds of dark gray dolomite at the top of the St. Charles formation," which are good stratigraphic markers, and for the first time recognizes the multiple lithology of the overlying Swan Peak beds.

In the Clarkston Mountain area (Loc. 8, fig. 1), originally mapped as undifferentiated Paleozoic by Butler et al. (1920, Pl. IV), the Garden City and Swan Peak formations are present, as they are in the southwest corner of the Randolph Quadrangle (Loc. 14, fig. 1), mapped in a similar fashion by Richardson (1941). The two formations also appear on reconnaissance maps prepared by the members of the Yale summer course in Geology during the 1946 and 1947 sessions in the southern portion of the Preston Quadrangle.

The writer is indebted to his wife, to J. Stewart Williams, to John Masters, and to William Swire for their assistance in the field; for their interest and suggestions concerning the study of the stratigraphy and fauna he is equally grateful to Prof. Carl O. Dunbar, James L. Wilson, C. R. Longwell, B. F. Howell, J. T. Gregory, K. M. Waage, H. L. Whittington, G. A. Cooper, W. A. Bell, and John Rodgers. The study was made possible by the Charles Schuchert Memorial Fellowship, and this paper is a part of a dissertation presented in partial fulfillment of the requirements for the degree of Doctor of Philosophy at Yale University.

STRATIGRAPHIC SUMMARY

Lithology and Thickness of the Garden City Formation. Lithologically the Garden City formation can be divided into two members, the lower two thirds being predominantly intraformational conglomerate and the upper third being very cherty. The lower member is a complex of interbedded layers of intraformational conglomerate, muddy limestone, crystalline limestone, and limy siltstone, with some layers of compact, thinly laminated, cryptocrystalline limestone. No sequence of beds that can be identified in more than one locality has yet been identified within this member; most of the beds appear to thicken, thin, lens-out, or change lithology within short distances. Only a few resistant, laminated, aphanitic limestones are at all persistent.

In addition to intraformational conglomerate, some of the coarsest of which is admirably exposed on a dip slope 87 feet above the base at Locality 11 (pl. 1, fig. 1), other characteristics suggesting a shallow water origin for the formation are channel fills (a good example 70 feet above the base at Locality 5), ripple marks (best exposed on a dip slope, 85 feet above the base at Locality 11; pl. 1, fig. 1), and cross-bedding in coarsely crystalline limestone (a good example at Locality 9; pl. 1, fig. 2).

Another feature of the lower member is the random scattering of white to light gray chert stringers in the lower 300 feet; these are best exposed at Locality 7, about 240 feet above the base of the formation, on several dip-slope surfaces, where they lie in rectilinear patterns strongly suggesting emplacement controlled by joints. These lower cherts are not to be confused with the black chert of the upper member from which they are quite distinct.

In the middle third of the formation the percentage of intraformational conglomerate (pl. 2, fig. 1) is considerably higher than in the lower third, where shaly and muddy limestones comprise much of the section. As in the lower third, individual beds not only lens-out, but their lithologies may change within several feet. In this portion there are in most localities several beds of a more persistent nature, composed of crudely laminated, crypto-crystalline, resistant limestone (pl. 2, fig. 2).

The upper one-third of the Garden City formation, the upper member, is sufficiently uniform throughout the area studied to be recognized with little more than a glance. Unlike the seemingly endless alternations and variations of lithologies of the lower two-thirds, this upper member is made up predominantly of thick, generally resistant beds of crypto-crystalline, dark limestone, composed of thin, discontinuous, very irregular laminae, one-half to one inch thick. The laminae are separated in most cases by paper-thin partings of mud, which give weathered vertical surfaces a peculiar wavy appearance (pl. 2, fig. 2; note uppermost layer above black chert, pl. 3, fig. 1).

The most striking feature of the upper member is its high chert content. The chert is black and in the form of nodules, stringers, and "discontinuous beds," oriented roughly parallel to the laminae of the crypto-crystalline limestones (pl. 3, figs. 1 and 2). In all the studied localities where the upper member is present there is one zone in which chert makes up approximately 50 percent of the exposed rock; above it the percentage slowly dwindles, until at the very highest limestone or dolomitic exposure very little or no chert is present. That this upper cherty zone has stratigraphic significance appears certain, not only because of its areal extent, but also because of its contained fauna.

Some rather coarsely crystalline dolomite is found at the very top of the formation at most of the studied localities and has been recognized by Duncan (personal communication) at Blacksmith Fork and St. Charles Canyon. That this dolomite is a *bona fide* sedimentary unit is open to question; at Localities 11 and 13 there is some suggestion that the replacement of limestone and dolomitic limestone by dolomite is secondary along faults or fracture zones. At Locality 1 there is undeniable evidence of the replacement of a bed of intraformational conglomerate by dolomite, but because of imperfect exposure it is not certain whether this is a case of penecontemporaneous or a much later secondary replacement. It should be noted that the irregular line separating the symbol for black chert and that of dolomite and dolomitic limestone in the diagrammatic sections (fig. 2) at Localities 13A and 11 represent irregular gradations, not unconformities.

At its type section (Loc. 1, fig. 1) the Garden City formation is 1,225 feet thick and is apparently little diminished or increased in a north-south direction. To the west, however, it attains 1,440 feet in Hillyard's Canyon (Loc. 5) and 1,760 feet at Clarkston Mountain (Loc. 8).

The Lower Boundary. In this study the base of the Garden City formation is placed below the lowest limestone bed which occurs above the thick, brownish-weathering, dark-gray dolomite at the top of the St. Charles formation (Williams, 1948, p. 1135). This lithologic definition of the lower limit of the formation appears to hold approximately the same position in a time sense throughout the area.

The Upper Boundary of the Garden City Formation. The upper boundary of the formation is not as easily determined as earlier descriptions imply; all of these, except Williams' state that the formation is overlain by the Swan Peak "quartzite." Actually, in neither the northwestern part of the Randolph Quadrangle (which includes the type sections of both formations) nor to the south is the relation so simple.

As recorded by Williams (1948, p. 1136) the Swan Peak formation is not entirely quartzitic. Its appearance at the type section (Loc. 1) is misleading, for the lower half of the formation is masked by rusty-colored, silty, sandy, and badly weathered quartzite rubble. At Locality 2, three miles to the west-northwest, there is an excellent exposure along the small gorge of Beaver Creek. Here one can see the thick massive, vitreous quartzites of the upper part overlying thinner, less firmly cemented, and more fossiliferous beds. Progressively lower in the section the quartzites thin, and more and thicker interbeds of shale are present. Limestone lenses occur in the lowest shales with only a few thin, calcareous sandstone layers. There is little doubt that the black shales exposed at Locality 2 are present but hidden by mantle and float at the type section, explaining the failure of previous workers to recognize the multiple lithology of the formation.

Southward the upper quartzites thin-out. At Locality 10 (fig. 8) they have decreased by half, and at Locality 12 (Blacksmith Fork) only thin interbeds of sandstone, shale, and lenticular limestones are present. At Locality 14A (Bucks Spring Road) the formation is entirely absent. To the west at Clarkston Mountain (Loc. 8) 50 feet of sandstones with calcareous cement directly overlie the Garden City formation. Although no shales were observed near the base by this author, a short interval overlying the sandstones and underlying the upper thick quartzites is covered in much the same manner as the lower part of the type section and may represent shale beds.

Since the contact is gradational from dolomitic limestone through shale to quartzite in some localities and since Ulrich and Cooper (1938) suggested in several instances that a part of one formation might be a facies equivalent of the other it was felt that the older unit might be thickening southeastward at the expense of the Swan Peak beds which disappear in that direction. There is, however, some faunal evidence which indicates that this is not the case.

The upper boundary of the Garden City formation is placed below the lowest shale, siltstone, sandstone, or quartzite bed of the Swan Peak sequence, although limestone lenses are encountered in some places above this horizon. This is the basis on which Williams (1948) mapped the formation in the Logan Quadrangle and the one found most satisfactory during this investigation.

FAUNAL ZONES AND CORRELATION

Although the fossils collected from the Garden City and Swan Peak beds include sponges, graptolites, pelmatozoans, brachiopods, gastropods, nautiloids, trilobites and ostracods, time has permitted a study of the trilobites only. A few brachiopods have been identified specifically, but most of them only generically. The graptolites have been compared cursorily with forms they resemble.

To date over 80 species of trilobites have been found in these two formations, all but two coming from the Garden City beds. In most cases the specimens are silicified and number over 25 per species. In a few instances forms are so distinctive that species have been based on much smaller samples.

As stated above, it appears probable that 12 *bona fide* faunal zones, designated by the letters "A" to "L" below, can be recognized within the Garden City formation; these are not founded entirely on trilobites. Further studies may show that too many zones have been set-up or that one or two should be based on different species. Since most of the trilobites are new, no attempt is made at this time to enumerate them in detail. The fauna of the Swan Peak beds is placed in zone "M."

Briefly, the zones characterized by previously described genera and species or by especially distinctive forms are set-forth below:

Swan Peak formation:

Zone "M"—

Eleutherocentrus petersoni Clark
Didymograptus cf. *bifidus* (Hall)
Orthis swanensis Ulrich and Cooper
Anomalorthis sp.

Numerous ostracoda, possibly referable to *Leperditia*

In addition this zone includes a distinctive, undescribed trilobite genus which has been collected also from the Upper Pogonip formation in Nevada and from the Lower Simpson group in Oklahoma; specimens from both are in the collections of the U.S. National Museum. Those from the Oklahoma area are in collections from U.S.G.S Localities 360K2 (labelled "Oil Creek formation") and 199t (labelled "Joins"). In both, the genus is associated with *Eleutherocentrus* and ostracoda which are very close to those found in the Swan Peak formation.

Garden City formation:

Zone "L"—

Anomalorthis sp.

Blastoidocrinus cf. *carchariaedens* Billings
Syntrophopsis transversa Ulrich and Cooper
Rhynchocamara n. sp.
Orthis sp. (close to *O. subalata* Ulrich and Cooper)

Zone "K"—

Blastoidocrinus cf. *carchariaedens* Billings
Nothorthis sp.
Diparelasma sp.
Hesperonomia sp.

Zone "J"—

Kirkella cf. *vigilans* (Whittington) (equalls so called *Asaphus curiosus* of other authors)
Eleutherocentrus n. sp.
Kawina n. sp.
Goniotelus ? sp.
Hesperonomia dinorthoides Ulrich and Cooper
Syntrophopsis cf. *polita* Ulrich and Cooper
Diparelasma sp.
Tritoechia n. sp.

Four other trilobite genera characterize this zone. One is an asaphid with a peculiarly fringed hypostome. The second is a proparian possessing a nasute anterior rim with a median pit between the rim and glabella. A Komaspid without glabellar furrows but with a pair of large nodes, one on either side of the glabella near the occipital furrow is the third, while the fourth is a very small *Dimeropyge*-like form.

Zone "I"—

Retiograptus sp.
Diparelasma cf. *typicum* Ulrich and Cooper
Hesperonomia sp.
Goniotelus sp.

This zone is based on collections from a single locality (8B); it may eventually be shown to belong in the base of zone "J."

Zone "H"—

Didymograptus cf. *nitidus* (Hall)
Dictyonema, 2 sps.
"*Xenostegium*" ? cf. *goniocerum* (Meek)

Zone "G"—

This is the most complex of the established zones. It is based on the occurrence of an *Apatokephalus*-like genus in which the anterior marginal furrow is tangent with the dorsal furrow in front of the glabella, pits are lacking in the marginal furrow, and each palpebral lobe possesses a curving furrow proximal to

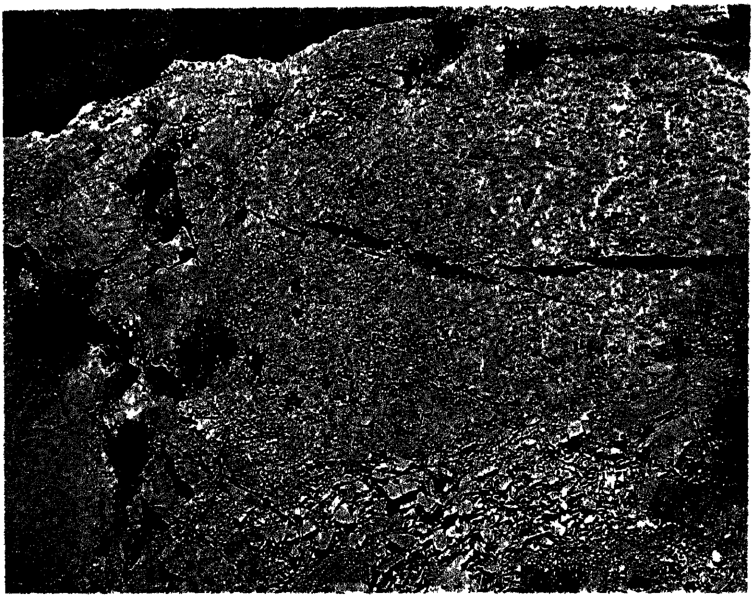


Fig 1—Coarse intraformational conglomerate bed overlying ripple-marked surface (left side) 87 feet above the base of the Garden City formation, Locality 11

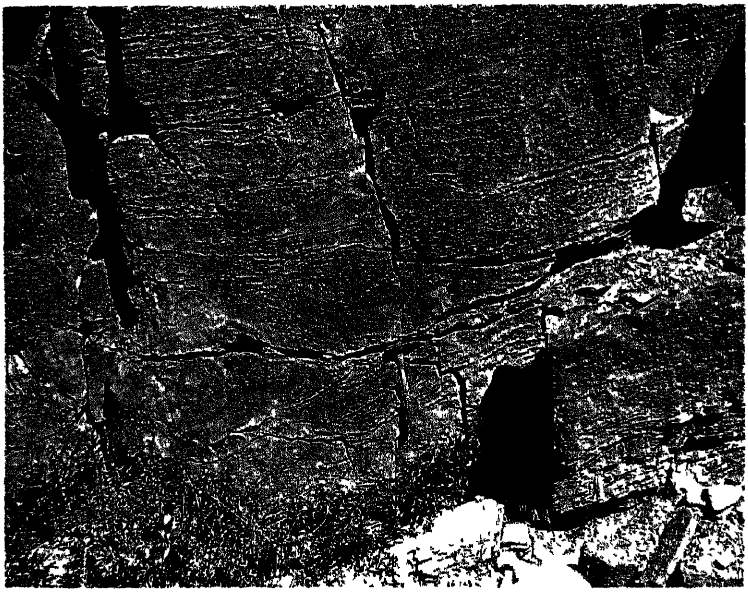


Fig 2—Cross-bedding in a coarsely crystalline limestone bed, approximately 500 feet below the top of the Garden City formation, Locality 9

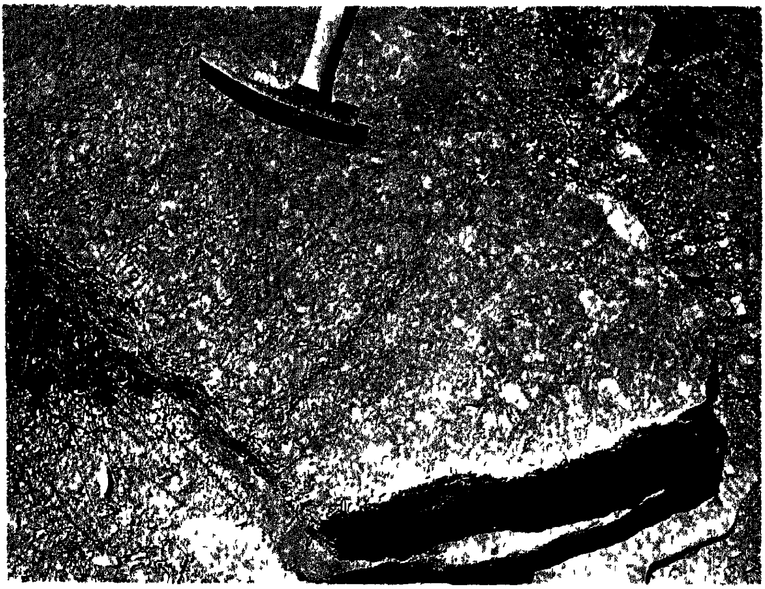


Fig. 1—A large float-block of intraformational conglomerate, 530 feet above the base of the Garden City formation, Locality 1

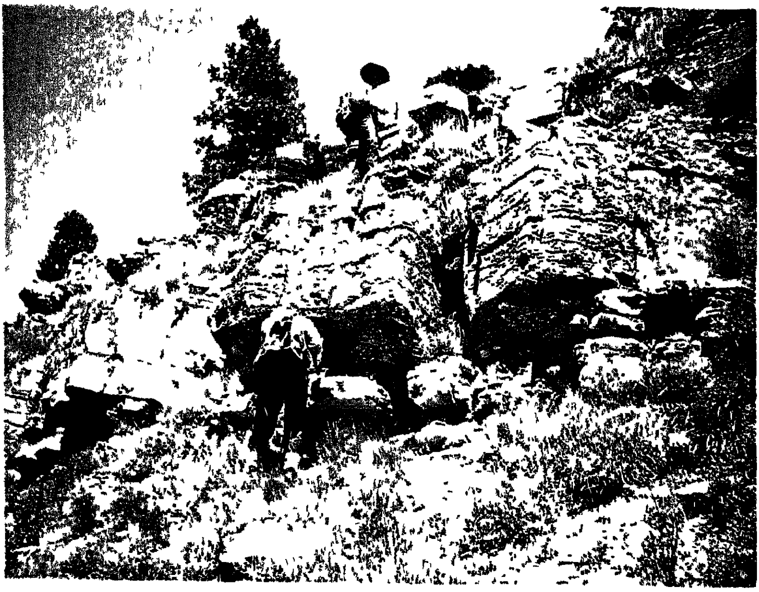


Fig 2—A thick resistant bed of crudely laminated, crypto-crystalline limestone, 215 feet above the base of the Garden City formation, Locality 5 Note weathered vertical surface

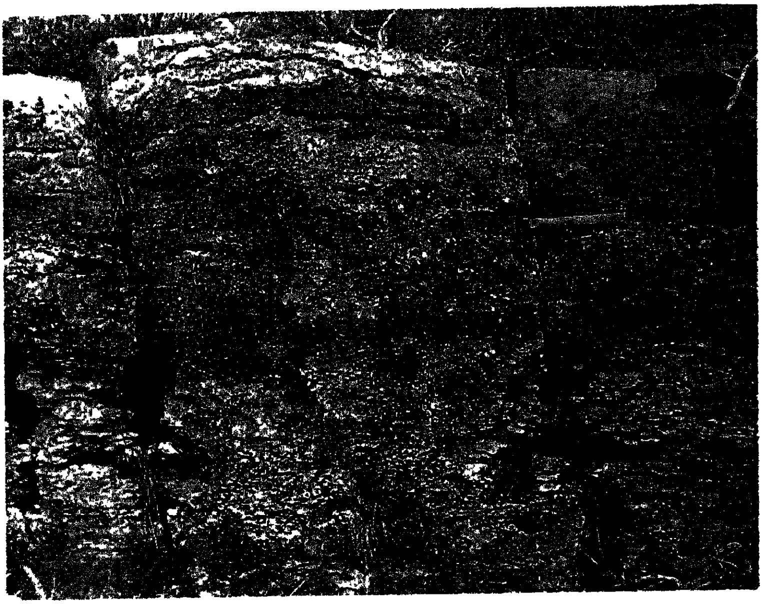


Fig 1—Irregularly laminated, crypto-crystalline limestone with "interbedded" discontinuous black chert "bands" approximately 200 feet below the top of the Garden City formation, Locality 3B

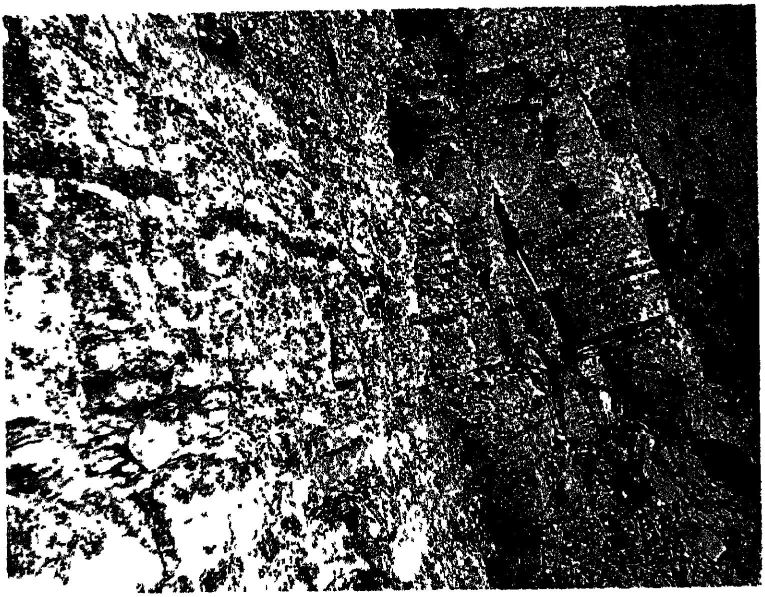


Fig 2—Black chert nodules, stringers, and "layers" in fine-grained dolomitic limestone, 200 feet below the top of the Garden City formation, Locality 8

and nearly parallel with the normal palpebral furrow. Within the range of this form occur two species of *Protophomerops,* on the basis of which an upper, "G(2)," and a lower, "G(1)," subzone are established. At the base of this zone a species is present which is apparently assignable to *Macropyge* Stubblefield. Near the top was found the only species assignable to *Jeffersonia.*

Zone "F"—

A total of 23 trilobite species have so far been recognized from this zone. These include four species of *Hystricurus* and one of *Protophomerops.* It may include the highest occurrence of the former and the lowest appearance of the latter, but such a possibility cannot yet be considered a fact.

Zone "E"—

Eleven species comprise this zone, including three of *Hystricurus*; one form is very close to *Beltella* Lake and another appears to be intermediate between *Pilekia* and *Protophomerops.*

Zone "D"—

Apheorthis cf. *meeki* Ulrich and Cooper
Leiostegium manitouensis Walcott

Zone "C"—

Nanorthis ? sp.
Syntrophina ? sp.
Symphisurina sp. (close to *S. spicata* Walcott)

Zone "B"—

Bellefontia sp
Clelandia n. sp.
Hystricurus, 3 n. sps.
Symphisurina sp.
Xenostegium sp.

This zone is also characterized by a small genus with a *Remopleurides*-like cranidium, but with a pygidium possessing four or five pairs of dagger-like spines.

Zone "A"—

Nanorthis sp.
Symphisurina sp.
Hystricurus sp.
Bellefontia ? sp.

This zone is possibly not a valid one. The trilobite species appear to be distinct from those of corresponding genera in the higher zones, but these differences may be due in part to different methods of preservation.

Although the above listing and comments are far from inclusive, they will at least serve as a generalized preview of the faunal zones, which are to be presented in detail with full systematic descriptions in a future publication.

With no more information than that presented above and with the assistance of the Correlation Chart for the Ordovician of North America (Ordovician Sub-Committee, Geological Society of America) now in preparation, it is possible to come to a few conclusions concerning the dating of the Swan Peak and Garden City formations. Richardson (1913, p. 409) originally considered the former to be of Chazyan age, this dating has been questioned more recently because of the occurrence of *Didymograptus* cf. *bifidus* in the shaly portions (Williams, 1948, p. 1137). There is in this author's opinion little doubt that the Swan Peak beds are at least as young as the Joins formation of Oklahoma (see discussion of zone "M" above); this is corroborated by the presence of *Anomalorthis* in zone "L" below the top of the Garden City formation; that genus is not known in beds older than Joins according to G. A. Cooper (personal communication).

Similarly, the species of faunal zone "L" may indicate that the uppermost portion of the Garden City formation is also of Joins age; in this connection it is noted that *Blastoidocrinus* has never been reported in beds older than Chazyan. It then appears probable that the Canadian-Chazyan boundary must fall within the upper Cherty member of the Garden City formation, somewhere between zones "L" and "J." The single species of *Jeffersonia* from the upper part of zone "G" is so close to *J. missouriensis* Cullison that it suggests the correlation of its subzone with the Rich Fountain formation of the Ozarks.

Since faunal zone "B" is a typical Tribes Hill-Stonehenge assemblage, it is concluded that the Garden City formation runs the full gamut of the Lower Ordovician and that no major change took place in the conditions of sedimentation in this area until after the beginning of Chazyan (Joins) time.

Furthermore the occurrence of species which are apparently very close to *Macropyge* Stubblefield and *Beltella* Lake may indicate contemporaneity of zones "E" to "G" with the Lower Ordovician of Britain.

LOCAL STRATIGRAPHIC SECTIONS

No attempt is made here to record the detailed lithologies of the 14 sections which were measured and from which collections were made; instead a very generalized account of each section is given below, noting the presence and relative position of any of the discovered faunal zones.

Supplementing the material which follows, seven of the sections are represented graphically in figure 2; as drawn they are very generalized since none of the more detailed lithology is indicated. Position of faunal zones is shown to the right of each of the sections. It is reiterated that the boundary drawn between cherty and non-cherty beds in sections at Localities 11 and 13A are not unconformities, but gradational contacts.

Locality 1. *Swan Peak and Garden City Canyon* (E. 1/8, Sec. 10, and N.W.1/4, Sec. 11, T. 14 N., R. 4 E.)

This is the type section of both the Swan Peak and Garden City formations. The Swan Peak formation is approximately 315 feet thick here; faunal zone "M" has been recognized in the interval 145-230 feet above its base. The lower 145 feet of the formation are masked by float.

The aggregate thickness of the Garden City formation at its type section is 1,225 feet, the upper 210 feet of which are assigned to the Cherty member. Only a single collection was secured from this member, 160 feet from the top of the formation; unfortunately silica replacement was so complete in the collected rock specimens that only a few fossils could be freed from the matrix. These belong to zone "J" or "K," but to which is not certain. Sponges are abundant in 18-foot intervals of massive, black, extremely aphanitic limestone 707-725 feet above the base of the formation, and large, orthoconic nautiloids four to five inches in diameter were found in aphanitic limestone 200 feet above the base. Faunal zone "B" was definitely found 50 feet above the base, but only a single species belonging to zone "A," a *Symphisurina*, was located approximately 20 feet above the top of the St. Charles formation

Locality 2. *East side of Beaver Creek, immediately south of the Idaho-Utah state line.* (E. 1/2, Sec. 35, T. 15 N., R 4 E.)

This locality was included to establish the presence of black shale beds in the lower part of the Swan Peak formation, which are not exposed at the type locality two and one-half miles to the east. Approximately 340 feet of the formation are exposed, of which the lower 145 feet are composed of black, flakey shale, with scattered limestone lenses near the base. Both the shales and the lower quartzite beds are abundantly fossiliferous; faunal zone "M" falls here.

Because of faulting only the uppermost beds of the Garden City formation are exposed at this locality; as a result no attempt was made at measurement.

Locality 3. *Fish Haven Canyon*

The Swan Peak and Garden City formations are exposed on both sides of the canyon approximately 4 miles west of Fish Haven, Idaho. The sections measured were not satisfactory, because of the thickness of plant and forest covering. In order to obtain any indication of thickness and lithologies it was necessary to run two traverses and then attempt to piece the results together; unfortunately there are large gaps in both resulting sections which prohibit perfect correlation between the two. The two traverses are indicated on the index map (fig. 1) and on the diagrammatic sections (fig. 2) as Locality 3A (E. ½, N.W. ¼, and W. ½, N.E. ¼, Sec. 17, T. 16 S., R. 43 E.) and Locality 3B (N ½, Sec. 7, T. 16 S., R. 43 E.). The exposed thickness of the Swan Peak strata are believed to approximate 535 feet and that of the Garden City formation 1,245 feet. The thickness of the upper Cherty member of the latter is between 230 and 360 feet.

At Locality 3B, in addition to the usual species found in faunal zone "M" in the Swan Peak beds a mold is tentatively identified as *Anomalorthis* sp. Zone "L" was located within 100 feet of the top of the Garden City formation at this same locality, but only a single representative, *Syntrophopsis transversa* Ulrich and Cooper, was found at Locality 3A, within 50 feet of the top. The interval from 240 to 285 feet below the top of the Garden City strata at Locality 3B contains a fauna assigned to zone "I," while forms indicative of zone "H" occur in the next lower 80-foot interval. No fossils were collected from lower beds.

Locality 4. *St. Charles Canyon* (6 miles west of St. Charles, Idaho, on west side of canyon; Sec. 13, T. 15 S., R. 43 E.)

The Swan Peak formation attains a minimum thickness of

500 feet at this locality. An apparent thickness of over 1,550 feet for the Garden City beds is attributed by Duncan (personal communication) to repetition by faulting, causing a seemingly considerable increase over that at Localities 1 and 3 to the south. Faunal zone "J" was located within 150 feet of the top of the formation and zone "B" within 100 feet of the base; no collections were made.

Locality 5. *East side of Hillyard's Canyon* (1.8 miles north of the head of the canyon; S.E. 1/4, Sec. 17, T. 15 S., R. 41 E.)

This locality and Locality 6 on the west side of the Canyon have provided the bulk of the trilobite faunas collected; both are more readily accessible than the type section of the Garden City formation and fossil preservation is far superior.

Although thinly represented 100 yards east of the top of the section by quartzite beds, only a thin veneer of rusty shale and silty "chips" of the Swan Peak formation overlies the Garden City beds at the measured traverse. Because of the thick vegetation separating the two points and the abundance of faults of large displacement along the canyon it is considered unwise to include any thickness for the younger formation here.

The thickness of the Garden City formation totals 1,440 feet of which the upper 370 feet belong in the Cherty member. The top 85 feet contain the fauna assigned to zone "L"; although zone "K" is present in the next lower beds the level of demarcation between it and zone "J" has not yet been satisfactorily established, probably because sampling was not definitive enough. The possibility also exists that zones "K" and "J" are intergradational. Both trilobites and brachiopods of zone "J" definitely occur throughout the lower 250 feet of the Cherty member.

Within the lower member of the formation faunal zone "A" was tentatively founded on a small collection from the base; zone "B" is present 75-85 feet, zone "C" is between 115 and 140 feet, and zone "D" appears in a thin interval 140 to 145 feet above the base. Only in a single one-foot muddy limestone bed, 305 feet above the base has the fauna of zone "E" been collected, while that of zone "F" has not been located to date in this section. Zone "G," with several of its subzones, appears from 428 to 920 feet above the bottom of the formation, and the interval from 945 to 1,020 feet contains faunal zone "H."

Locality 6. *Crest of Ridge on west side of Hillyard's Canyon*

This section was measured for supplementary use with Locality 5; only the lower 750 feet of the Garden City formation is well-exposed. From here a prolific silicified trilobite fauna has been secured, from the same zones as at Locality 5 through zone "G" with the exception of "A," "C," and "D." The forms of zone "E" are abundantly represented within a considerably thicker interval, and those of "F," which were not located at Locality 5, are present in profusion.

Locality 8. *West side of Clarkston Mountain* (Sec. 25, T. 14 N., R. 3 W.)

This is the most northwesterly of the sections in the area studied. The base of the Garden City formation is exposed approximately 2,500 feet above the level of U.S. Highway 191. The Fish Haven dolomite overlies the Swan Peak formation disconformably at this locality.

The thickness of the Swan Peak beds totals 570 feet, the lower 50 feet being composed of calcareous quartz sandstone. Although 1,764 feet of Garden City beds were measured, it is barely possible that there may be some minor repetition of beds through faulting; because faults of large displacement were recognized on either side of and above the block on which the measured traverse was run, special attention was given this possibility. It is believed that the measured thickness is original.

At this locality the upper Cherty member of the Garden City formation is about 400 feet thick. Within 155 feet of the top representatives of both faunal zones "L" ad "K" were found but the boundary between the two is not as yet definite. Both trilobites and brachiopods of zone "J" are abundant in a 10-foot interval 245 feet from the top of the formation.

At the top of the lower member, from 1,135 to 1,340 feet above the base of the formation, specimens were collected from zone "H." Only 913 feet above the base a single specimen of *Kirkella* Kobayashi was secured. In the interval from 330 to 615 feet above the base a few forms assigned to zones "C" and "D" were located, but the two do not appear to be separable as easily as at other localities. Trilobites of fauna "B" occur in scattered beds from 20 to 295 feet above the St. Charles formation.

Locality 9. *Twin Bridges Dugway, Logan Canyon* (N.W. ¼, S.W. ¼, Sec. 34, T. 13 N., R. 3 E.)

This locality is situated in the abandoned road-cut on the west side of U.S. Highway 89 at the north end of the Twin Bridges dugway. It was visited exclusively for the collection of graptolites. The exact stratigraphic position of the zone cannot be ascertained but it is estimated to be 500-600 feet below the top of the Garden City formation. Graptolites include two species tentatively assigned to *Dictyonema* and one species of *Tetragraptus* close to *T. quadribrachiatus.* A few feet above the graptolite beds a massive limestone strata contains abundant silicified sponges possibly referable to *Receptaculites.*

Locality 10. *Junction of the Logan River and its Right Fork* (N.E. ¼, Sec. 18, T. 12 N., R. 3 E.)

The Swan Peak formation and the upper 150-200 feet of the Garden City beds are exposed on the north side of the Right Fork; on the west side of U.S. Highway 89 the upper Swan Peak beds are well displayed.

Here the Swan Peak formation is 240 feet thick, a considerable decrease from its total to the north. Collections made from the top 100 feet of the Garden City strata were unfortunately lumped together; it is now evident that faunal zone "L" and zone "K" are both present in the interval.

Locality 11. *Green Canyon* (N. E. ¼, Sec. 18, and N. W. ¼, Sec. 19, T. 12 N., R 2 E.)

The complete St. Charles, Garden City, Swan Peak, Fish Haven sequence is exposed on both sides of the Canyon which cuts approximately at right angles across the strike of the beds. The lowest beds of the Garden City formation are found approximately 1,600 yards east of the Canyon's mouth.

Williams (1948, p. 1136) gives a thickness of 339 feet for the Swan Peak beds at this locality, a figure very slightly in excess of this author's measurement. The fauna of zone "M" is abundant in silty and sandy layers of the lower 180 feet and in limestone lenses interspersed throughout the lowest black shales.

The Garden City formation attains a thickness of 1,405 feet, of which the upper 365 feet belong to the Cherty member. Faunal zone "L" occurs within 50 feet of the top; zone "B" is approximately 125 feet above the base of the formation, and one of the sub-zones of zone "G" has been located 490 feet above the base.

Locality 12. *Blacksmith Fork, north side of Canyon* (S. ½, Sec. 2, T. 10 N., R. 2 E.; near power house)

The thickness of the Swan Peak formation here is considerably diminished to 105 feet, and none of the thick quartzites of the type section are present. None of the beds is over one foot thick; in general three- to six-inch quartzite and sandstone layers, two- to four-inch greenish shale layers, and one- to three-inch limestone lenses are interbedded. There are nearly as many limestone lenses near the top as quartzite and sandstone layers near the base. The formation has little surface expression, because of the unresistant nature of the shales and the silty character of the quartzites and sandstones.

The Garden City formation has been measured at this locality by Duncan, who found it to be 1,160 feet thick (personal communication). Although his data are in my possession, the opportunity has not arisen to study his fossil collections at Princeton University. Since Duncan's collecting did not include the upper 100 feet of the formation, this author secured specimens which have proved to be identical with those collected from the same interval at Locality 10.

Locality 13. *Round Hill, north of Mantua, Utah* (S.E. ¼, Sec. 10, and N. ½, Sec. 15, T. 9 N., R. 1 W.)

At this, the most southwesterly of the studied sections, the Swan Peak formation is 245 feet thick; massive quartzites comprise the upper 55 feet, thin beds of reddish quartzites and siltstones the next lower 30 feet, and olive-green shales with scattered sandstone layers the lower 160 feet. No better locality was found for the collection of species belonging to faunal zone "M," especially of *Didymograptus* cf. *bifidus*.

The Garden City formation totals 1,390 feet in thickness, with the upper Cherty member consuming the top 425 feet. Although faunal zones "L" and "K" were not located, beautifully silicified trilobites and brachiopods of fauna "J" were collected in the interval from 270 to 355 feet from the top of the formation; this collection also includes a few fragmentary specimens believed to be *Receptaculites*.

feet, of which the upper 365 feet belong to the Cherty member. Faunal zone "L" occurs within 50 feet of the top; zone "B" is approximately 125 feet above the base of the f

Collections made to date from the lower beds of the formation are not satisfactory for ac ate correlation with other

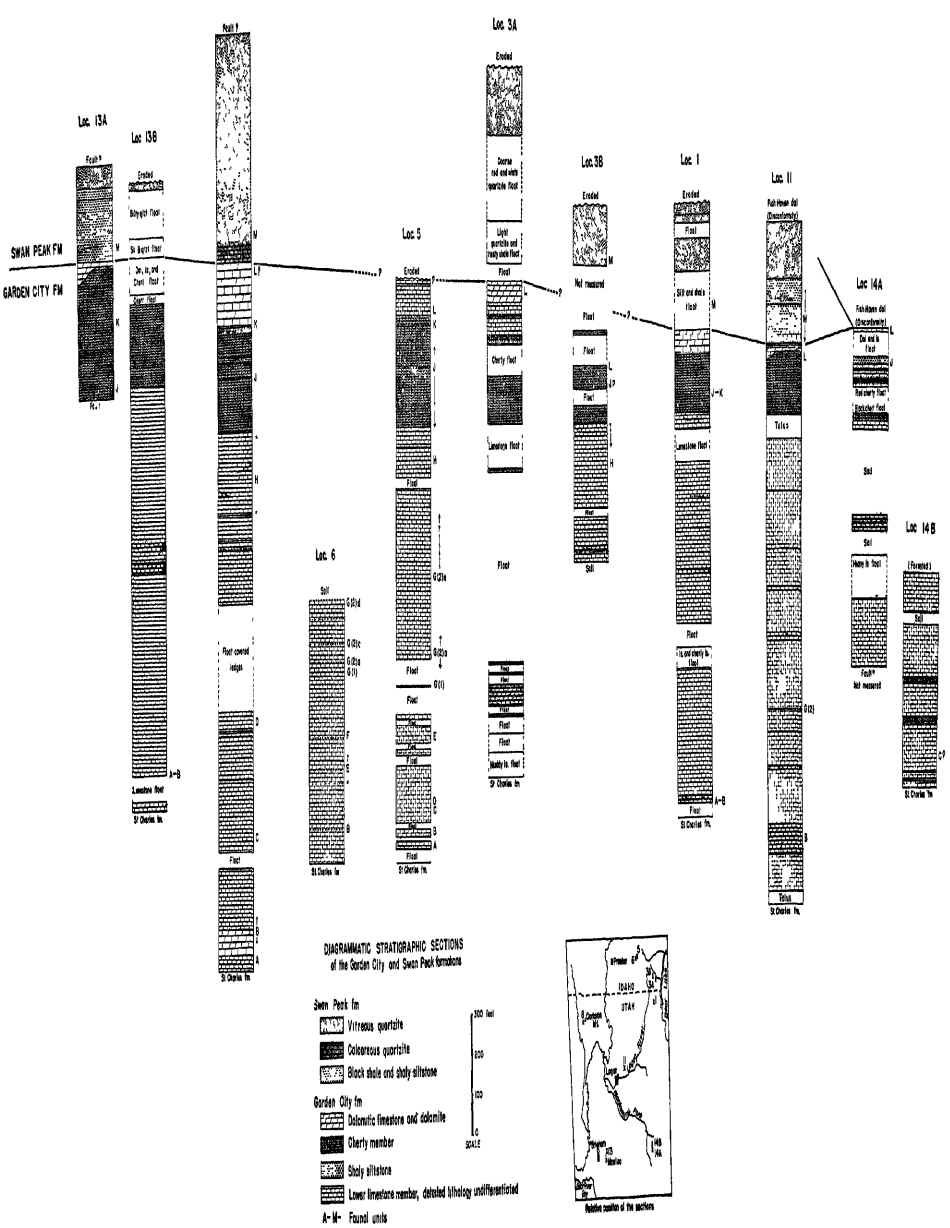


Figure 2 Generalized stratigraphic sections of the Garden City and Swan Peak formations. Refer to figure 1 for locations Note contact between dolomite and top of cherty beds in sections 13A and 11 is gradational, not unconformable

Collections made to date from the lower beds of the formation are not satisfactory for accurate correlation with other sections. Two poorly preserved trilobites were secured 230 feet above the base which are believed to belong in zone "E"; the forms taken from 90 feet above the bottom of the formation include *Hystricurus* cf. *cordai*, not found in any of the other sections, and a few specimens of *Nanorthis* and *Syntrophina*(?), which are considered to be the same as those from zone "C" at Locality 5.

Locality 14. *Davenport Hollow, north of Monte Cristo Ranger Station* (Between S.E. 1/4, Sec. 5, and W. 1/2, Sec. 17, T. 9 N., R. 4 E.)

In almost every respect this is the least satisfactory of all the localities visited. Because of heavy vegetation and complex faulting no single section was found in which the entire Garden City formation could be measured. To tie-in the bottom and top of the section an attempt was made to trace a key bed for more than a mile between two traverses, resulting in a computed thickness of 1,185 feet for the formation; this figure is not considered reliable, but roughly indicates that the formation has changed little in thickness southward from its type section.

This locality's main point of interest is the absence of the Swan Peak formation. It was at first believed that a 20-foot interval of light brown, flakey shale and silty shale, which lies 70 feet below the Fish Haven dolomite, was equivalent to some of the lower shales of the Swan Peak. However, between these shales and the basal Fish Haven beds there is typical cherty limestone of the type found in the upper part of the Garden City formation, topped by four feet of brownish arenaceous dolomite. In this dolomite occurs a fauna which undeniably belongs to zone "L." Furthermore, the shales below contain a very distinctive proparian trilobite species which is especially characteristic of zone "J."

Within a 30 foot interval approximately 90 feet above the base of the formation specimens of two species of *Hystricurus*, one species of *Symphisurina*, and one species of *Syntrophina* were collected; little more can be told from these than that they lie below the upper limit of zone "F."

Interpretation of the Local Stratigraphy. During Early Ordovician and part of Chazyan time the Logan Quadrangle and its environs were the locus of deposition of a great thickness of interbedded and interlensed calcirudites, calcarenites, and calcilutities, composing the Garden City formation. Intraformational conglomerates, ripple marks, and cross-bedding in the limestones of the formation attest to its shallow-water origin. Toward the close of the Early Ordovician large amounts of silica were apparently supplied to the area, leading to the formation of an upper Cherty member.

Although the Garden City beds are some 500 feet thicker on the west at Clarkston Mountain than at the type area, there is in general relatively little change in the vertical dimension throughout the studied area. To account for the eastward decrease in thickness the possibility was at first contemplated that the oldest beds present at Clarkston Mountain (Loc. 8) might have wedged- or lapped-out against the eastern side of the Cordilleran geosyncline; because the same faunal units are found near the base of the formation on both east and west it is concluded that the difference must be attributed to a greater rate of deposition nearer the center of the trough. This is further substantiated by the increased spread between faunal zones in the western sections.

In the Swan Peak formation there is evidence for a marked change in these conditions early in Chazyan time, a change leading to the deposition of dark shales, shaly sandstone, and finally clean, white quartzites. Limestone lenses within the shales of the more easterly sections indicate that there were occasional relapses to the Garden City type of sedimentation and that the change was probably gradational to some extent.

Of great interest is the fact that the Swan Peak beds thin and disappear to the southeast within the Logan quadrangle with the shales persisting furthest to the east. The formation may have existed as a uniform sheet over the entire area at one time; since there is an unconformity representing all the rest of Middle Ordovician time between it and the Richmondian Fish Haven dolomite, it is conceivable that the southeast portion of the region may have been broadly upwarped and the Swan Peak beds bevelled by erosion to leave a northwestward thickening wedge. An alternate hypothesis calls for a northwesterly source of the sediments with the shales and a few of the sands being swept as far to the southeast as Blacksmith Fork (Loc. 12), but with the bulk of the coarser arenites being deposited closer to the source.

If the second of these hypotheses is correct, one could expect to find each of the faunal zones progressively higher relative to the Garden City-Swan Peak formational boundary in a northwesterly direction. The answer can undoubtedly be found in the Pogonip limestone and Eureka quartzite in western Utah and Nevada.

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GROWTH PATTERNS IN THE CERATOPSIA

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ABSTRACT. The deformed coordinate system of D'Arcy Thompson is applied to both ontogenetic and phylogenetic series of ceratopsian dinosaurs. The results confirm the phylogeny of Lull, 1933, and the relative growth studies of Gray, 1946. An evaluation of the method and its relation to the relative growth formula is provided.

SIR D'Arcy Thompson in his book *On Growth and Form* showed, that, as a one to one relationship exists between homologous parts of the bodies of related animals, this relationship could be expressed in terms of a system of coördinate lines undergoing successive deformation. He suggested that this interpretation of relationship might provide a method for predicting unknown intermediate stages in the evolution of animal groups. He, himself, applied such a system to the phylogeny of the horse, comparing the actually known stages with those derived by deformed coördinates from beginning (*Hyracotherium*) and end (*Equus*) forms only. The agreement was very good. Colbert, 1935, used the method to show form changes but did not attempt to construct a phylogenetic system. Richards and Riley, 1937, used a modification of the method to show the differences between normal and abnormal growth in amphibian larvae. The method has been discussed in general by Woodger, 1945.

THE CERATOPSIA

The present study attempts to apply the deformed coördinate method to the phylogeny of the ceratopsian dinosaurs, a group of ornithischian reptiles from the Cretaceous. Growth of this group was previously studied by means of Huxley's relative growth equation by Gray, 1946, and it therefore seemed to be good material for testing the usefulness of the Thompsonian coördinates.

The specimens used were the series of skulls figured in Hatcher, Marsh, and Lull, 1907, and Lull, 1933. A few other sources supplied single specimens. Enlarged drawings were made from photographs, and coordinate lines were drawn directly upon them. The base grid was chosen arbitrarily for convenience although an attempt was made to have the evenly spaced lines cross as many definite landmarks as possible. Neither brow nor nasal horns were considered in the system as they are too frequently broken or distorted.

The first base selected was the Asiatic *Protoceratops andrewsi* described by Brown and Schlaikjer, 1942, because

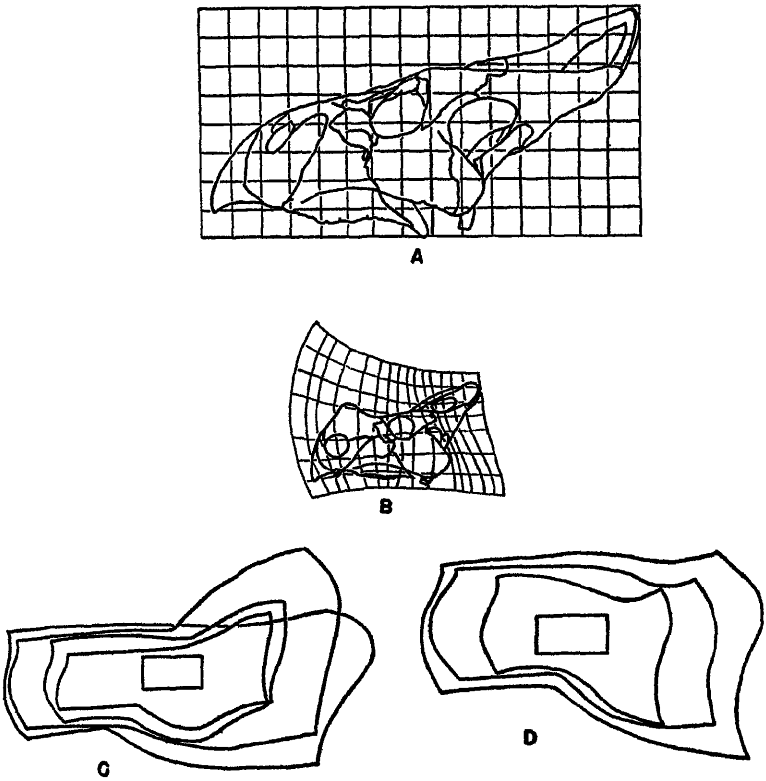

Figure 1. A *Protoceratops andrewsi* with the basic rectilinear coordinate system B. Same coordinate system deformed to fit *Leptoceratops cerorhynchus.* C Same coordinate system (outlines only) for *P andrewsi* (innermost), *Chasmosaurus belli, C. brevirostris, Pentaceratops sternbergii* and *Torosaurus gladius* (outermost). D. Same coordinate system (outlines only) for *P. andrewsi* (innermost), *Monoclonius flexus, Triceratops prorsus,* and *T. horridus* (outermost).

of its small size, moderate crest development and absence of horns (fig. 1A). Previous studies of the ceratopsian material, Gray, 1946, had shown that there was a discontinuity between the relative dimensions of *Protoceratops* and those of the remainder of the group (fig. 1). This discontinuity resolved itself into a nearly constant deformation of the rectilinear coordinates in all of the higher forms when *Protoceratops* was used as a base. There is a downward displacement of the post-orbital region superficially apparent from the consistently inferior location of the temporal opening. This displacement is more marked in the ventral than in the dorsal contour lines.

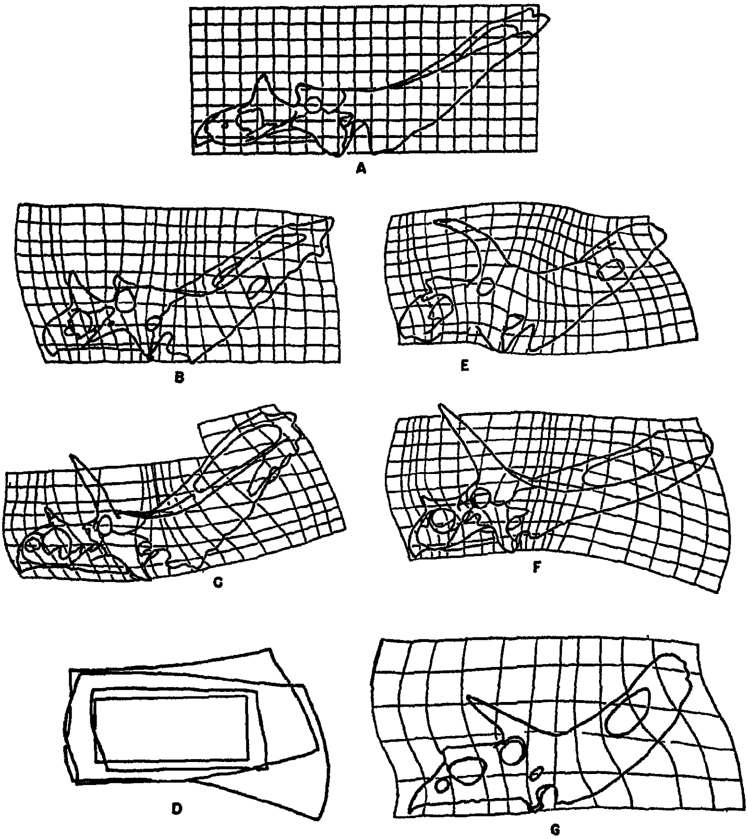

Figure 2 A. *Chasmosaurus belli* with basic rectilinear coördinate system. B. Same system deformed to fit *C. brevirostris*, C. *P. sternbergii*, D. Superimposed outlines of A, B, C, E and F, E. *Arrhinoceratops brachyops*, F. *Torosaurus gladius*, G. *T. latus*.

While it is slightly more prominent in the *Monoclonius-Triceratops* series (the "short-crested" line of Lull) it is present in all of the forms examined, showing neither a constant increase nor decrease as one goes up the series (figs. 1C and D). It is, however, a smooth deformation and *may be drawn with unbroken lines.*

Leptoceratops cerorhynchus (AM 5464), as reconstructed by Brown and Schlaikjer, does not show any trace of this deformation found in the higher forms but instead shows a quite different pattern of its own (fig. 1B). The extent of reconstruction renders judgment difficult but it does not seem to lie between *Protoceratops* and the other Ceratopsia.

This constant deformation of the Ceratopsia with respect to the base, *Protoceratops*, obscures other changes which may have taken place. New series were therefore constructed. For the "long-crested" line of Lull, *Chasmosaurus belli* (AMNH 5402) was selected as the base form (fig. 2). This skull, from the Belly River formation of the upper Cretaceous has been considered the earliest member of the series culminating in the giant *Torosaurus* of the Lance formation. The only marked deformation in the series is in the angle which the crest makes with the rest of the skull. Both *T. gladius* and *T. latus* from Yale show considerable downward bending of the crest but it should be noted that the skull from the Hell Creek formation of South Dakota assigned to *T. latus* by Colbert, 1947, shows no such deformation with respect to *Chasmosaurus belli* (fig. 2G). It is probable that the crest was easily distorted during preservation and not impossible that its angle with the remainder of the skull was somewhat variable from one individual to another during life. Neglecting this angular deformation of the crest, the phylogenetic growth in the *Chasmosaurus-Torosaurus* line, is very nearly uniform, with a slightly accelerated increase in the posterior lateral edges of the crest.

Arrhinoceratops brachyops (*ROM* 5135), placed doubtfully between the long- and short-crested lines by Lull, 1933, has been shown to belong to the long-crested series on the basis of relative dimensions and the deformed coordinates confirm this conclusion. The deformation with respect to *Chasmosaurus* indicates that, while it is closer to this line than to the *Monoclonius-Triceratops* series, it shows a certain downward bending in the post-orbital region which is reminiscent of that seen in all of the forms when deformed to the *Protoceratops* base.

In the short-crested line, *Monoclonius flexus* (YPM 2015) provided the base for the coordinate system (fig. 3). In general, even less distortion appeared than in the *Chasmosaurus-Torosaurus* series. There was, however, much more trace of the deformation seen with respect to *Protoceratops*. In other words, the downward bending of the post-orbital region noted between *P. andrewsi* and *M. flexus* was still continuing in the higher forms although to a much smaller extent.

Both this study and the previous one with relative growth seem to indicate that no better classification of the group can be made than that of Lull in 1933.

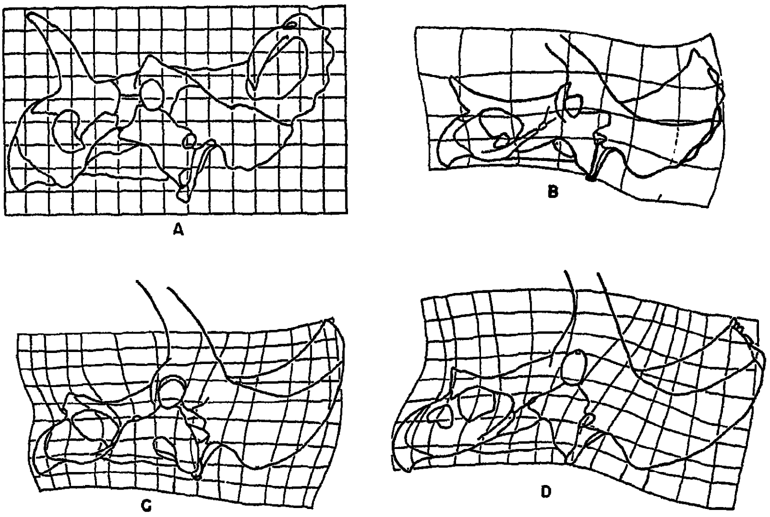

Figure 3. A. *Monoclonius flexus* with basic coordinate system. B. same system deformed to fit *Triceratops prorsus*, C. *T. elatus*, D. *T. calicornis*.

An attempt was made to follow the deformation of both phyletic lines backward in time by extrapolation from *T. prorsus* through *M. flexus* (fig. 4A). It was found impossible to continue the change at the same rate to a point as far "below" *M. flexus* as *T. prorsus* is "above" it. Before reaching such a stage, parts of the deformation become geometrically impossible, that is, the figure begins to buckle and fold. It is possible to reach a stage about three quarters of the way and this is shown in fig. 4A. This completely hypothetical skull shows some resemblance to *Protoceratops* in its small crest and high nasal region, but the position of the temporal opening in relation to the orbit is like that in the higher forms. This character shows no sign of approaching the protoceratopsian condition.

The similar extrapolation from *Pentaceratops sternbergii* through *Chasmosaurus belli* does not at all resemble either of the protoceratopsian skulls. It, in fact, looks like nothing more than a smaller specimen of *Chasmosaurus*. There is no trace of a tendency to approach *Protoceratops* in the long-crested line.

Changes in the ontogenetic series of skulls of *Protoceratops andrewsi* from Brown and Schlaikjer, 1940, were examined from a series of deformed coördinates based upon the youngest specimen (AM 6419) (fig. 5). A series of five skulls shows an excellent set of progressive deformation culminating in the "Old Male" (AM 6432). The greatest change is in the great increase in the depth of the nasal region with age. This is not similar to the condition in the higher Ceratopsia where the nasal region remains relatively undisturbed, all of the growth being localized in an actual horn instead of the more diffuse involvement of the surrounding bone.

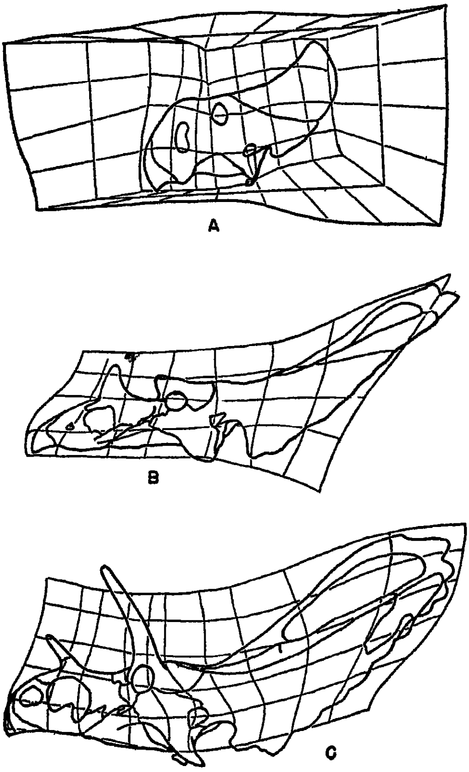

Figure 4 A. Hypothetical skull resulting from extrapolating coordinate system based on *Monoclonius* from *T prorsus* (outermost) through *M flexus* backward in phylogenetic time (see Text) B. coordinate system based on *M flexus* deformed to fit *C. belli.* C. coordinate system based on *T prorsus* deformed to fit *P. sternbergii.*

In no way does the development of the young *Protoceratops* resemble the phylogenetic changes of the sub-order as a whole. There is no tendency for the temporal opening to shift toward the position found in the higher forms and the increase in the size of the crest hardly forecasts the great crests of the future. The specimens tentatively called females by Brown and Schlaikjer proved to fit the deformation at all points but the nasal region. There is no doubt that these skulls are slightly but consistently different from the other five and this difference may well represent sexual dimorphism (fig. 5A).

DISCUSSION

The Thompsonian grid, being constructed so that lines pass through homologous points in a series of skulls, it follows that

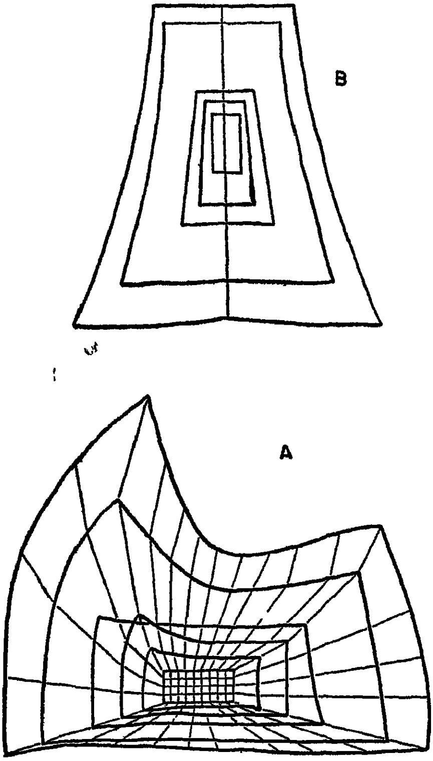

Figure 5.—A. Ontogenetic series of outlines of successive deformations of coordinate system taking place in the growth of *Protoceratops andrewsi.* The grid is based on the youngest specimen (AM 6419). The fourth outline from the inside is that of a putative female B. Similar outlines for dorsal aspect of skull of *P. andrewsi.*

the distance between two points of intersection on one skull is homologous with the distance between the equivalent two points on another skull in the same series. It has been shown in a previous paper (Gray, 1946) that many dimensions on the ceratopsian skulls follow the Huxley equation $y = bx^k$ in both phylogeny and ontogeny. It should follow from this that any pair of line segments in the Thompsonian grid will be similarly related to the homologous pairs of line segments of the same grid distorted to pass through homologous points of another skull of the series. If this "distortion" is simple, uniform growth without bending, and the line segments remain straight, the resulting patterns are capable of ready analysis. If the growth rate in all directions be equal, and the initial dimensions be equal, then b in the Huxley formula equals 1.0 and K equals 1.0. The resulting Thompsonian series will have straight lines connecting all equivalent points of the series (fig. 6D). If b remains equal to one and K becomes greater than one the midpoints of the line segments alone will lie in a straight line and the remaining equivalent points will fall on logarithmic curves concave toward the segment chosen as x. Should K become less than unity the curves will become concave toward the segment chosen as y. This expresses the fact that the value of K becomes its reciprocal if the dimensions chosen as x and y are interchanged (fig. 6C and E).

But such simple growth systems as these are not common, and in practice, in the Ceratopsia, one finds that the axes are deformed and that many varying rates of growth are involved so that rarely do the equivalent points lie in a straight line. In addition to this, the skull is not homogeneous and will resist deformation more in one direction than in another and thus the increase in length of a region due to growth may appear, chiefly, at that end least firmly attached to more slowly growing structures When, in addition to these complications, it is considered that certain growth rates may change during the phylogeny of a group while others remain constant, it is impossible accurately to analyze the deformed series into its factors without a nearly perfect set of specimens. For this reason it has seemed best to treat most of these lines as being straight. Similarly, while it is possible to analyze the patterns of angular change in the direction of a line in the Thompsonian grid, there is insufficient material to be sure that the skull actually obeys such laws as this analysis might indicate.

When progressive deformations are carried far enough there usually appears a point at which further deformation is no longer possible on a plane surface. Simply expressed, the condition is usually that in which three points form a triangle, one of whose sides grows fast enough to exceed the total of the other two. Such a condition is geometrically impossible and biologically absurd. It was reached during the attempt to extrapolate backward in time in both the *Monoclonius-Triceratops* line and the *Chasmosaurus-Pentaceratops* series. Long before such a condition occurred in nature the forces involved would produce a bending effect upon some one of the structural members involved. An indication that this is already happening in the "Old Male" *Protoceratops* appears in figure 5A. The slow growth of the frontal bones above the orbit with the great increase of the vertical height in the nasal region has produced the elevated nasal prominence. Nevertheless, this growth has placed a strain on the lower portion of the skull so that there is a tendency for it to give in a downward direction and thus produce a more hooked beak. To what extent such rapid growth of one region may affect the history of a group such as the ceratopsia is not clear. They remained well within the bounds of geometrical possibility but had

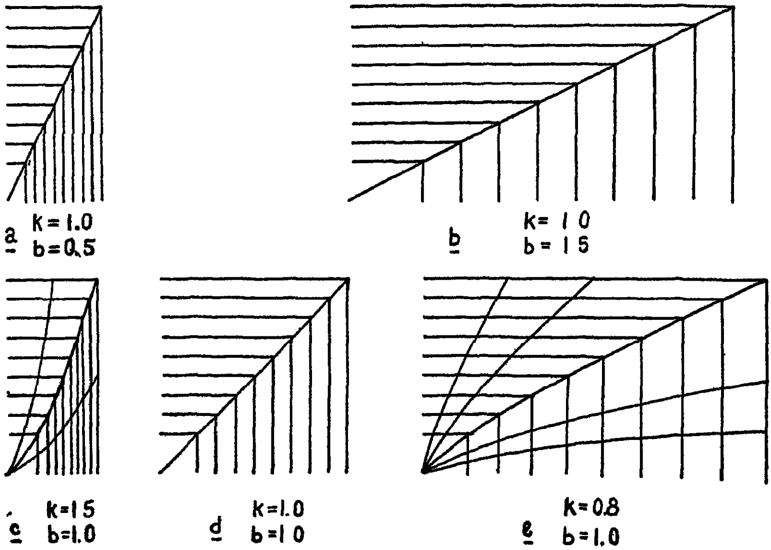

Figure 6. Appearance of quadrants of deformed coordinate systems showing the effects of varying *b* and *k* in the Huxley equation.

they not become extinct, and had they continued to increase in size with their same relative skull growth rates they would have reached this interesting stage. As Simpson, 1944, has pointed out, this would affect only the oldest animals first and, even though the resulting stresses might prove fatal, through warping of the upper jaw with resulting mal-occlusion, for example, it would exercise no selective effect until it had progressed far enough to make itself felt in animals of breeding age. Thus, the impending phylogenetic crisis might be first apparent in a decreasing individual life-span.

From these considerations it would appear that there are real limitations to the extent with which orthogenetic or "straight-line" evolution may operate. An increasing dimension need not reach the bizarre and inefficient size necessary to become an impossible burden upon the animal but need only grow to such a point that it can no longer be fitted with other, more slowly growing dimensions. This might well occur in a structure of no great absolute size.

Orthogenesis can operate, then, only within the framework of a Thompsonian grid having line segments of positive sign. Overgrowth of one region may imply that a distant dimension become negative ("folding" or "buckling"), and, while the other side of zero holds no terrors for the mathematician, it is a land closed to dinosaurs.

In summary it may be said that the deformed coördinate system should, under careful analysis, yield, at one stroke, the same information to be obtained from a large series of relative growth measurements. The relationship of ontogeny to phylogeny within such a group as the Ceratopsia may be illustrated both in terms of relative growth of single dimensions and the deformed coördinates in figures 7 and 8. In practice, the required accuracy in the identification of homologous points is not easily achieved and the use of the system is thus markedly limited. It should also serve to indicate probable lines of evolution within the series, and, indeed, will no doubt succeed where more perfect material is available. At present, it seems to be rather less useful, despite the striking nature of its results, than the method of comparing single pairs of measurements which may be available even when the material is fragmentary. With either method it is possible to produce totally spurious relationships; for two points may always be connected by a straight line although evolution may have proceeded quite differently. Thus, there is a perfectly possible transformation between *Protoceratops* and *Monoclonius* which may or may not be the actual change which took place. There is a deformation of coördinates based upon *Monoclonius* and applied to *Torosaurus* which represents a change which did not occur at all, yet, in neither case would the data alone give one the clue to this. Both methods tell us only that in the case

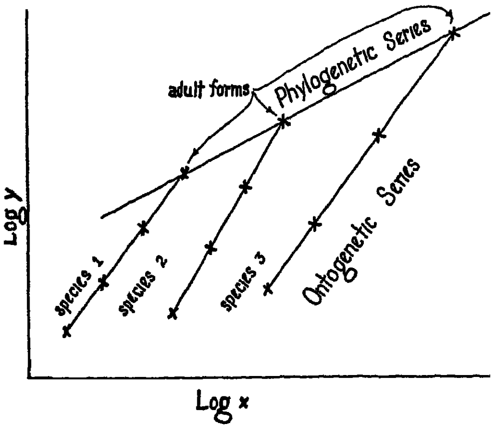

Figure 7. Generalized relation between ontogenetic and phylogenetic relative growth. Where two dimensions, x and y, are compared, each species has its own growth curve with values of *b* and *k* peculiar to it. The phylogenetic curve of the *adult* forms intersects each ontogenetic curve with independent values for the variables

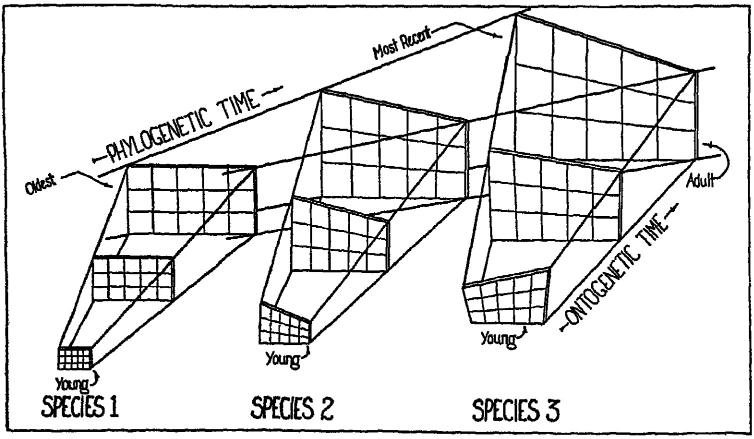

Figure 8 Generalized relation between deformed coordinate systems for three species. As in Figure 7, each ontogenetic series culminates in an adult form which falls into line with other adult forms to create the phylogenetic series.

of two different forms there are certain differences and, *if* the forms are related, it is in a certain way. In spite of these drawbacks, the analysis of the Ceratopsia confirms the relationship previously suggested and further emphasizes the great gap between the Protoceratopsidae and the Ceratopsidae.

Even a casual glance at the deformations produced by this method shows that boundaries between bony elements of the skull are in no sense boundaries of growing systems. The crest behaves as a unit without reference to the boundary between the parietals and squamosals. The orbital region is not the result of a ring of bones growing at different rates but acts as if the whole region were under a single control. Even such units as the crest and the orbital region are so smoothly integrated with one another that one feels that the true "field" can be no less than the entire skull. It is possible that this integration can be broken down into growth centers of some sort, each under some overall control by the entire animal.

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OBITUARY

DR. MAX BLANCKENHORN (1861-1947)

Max Ludwig Paul Blanckenhorn, pioneer German geologist whose contributions to the advancement of African and European geology are notable, died at his home in Marburg an der Lahn, Germany, in January 1947.

He was born on April 16, 1861, in Siegen where he had his education and was graduated in 1884. He was attracted to the Middle East by reading the scriptures, so that when he went to study further in Strassbourg, he developed his idea of working out the whole African geology. In 1887 he went to Berlin and carried out research work under Professors von Richthofen, Kiepert, and von Bezold. In 1888 he travelled with von Luschon and then alone to Syria and the Euphrates. He received his Doctorate Degree in 1891, his thesis being "Geology of Africa." In 1894, he travelled to Palestine, Sinai and Egypt, and in 1897 he was appointed as a field geologist and later as a paleontologist in the survey of Egypt on the strength of Zittel's recommendation. He mapped large parts of Egypt, travelled the whole of the Middle East, and contributed to the geology and paleontology of that area. In 1898 he became acquainted with Schweinfurth (1836-1925), an acquaintance which later developed into great friendship. They corresponded with each other until Schweinfurth's death, a correspondence which is partly deposited in the *Geologen Archiv.* Blanckenhorn then accompanied Stromer von Reichenbach on his famous expedition to Egypt collecting vertebrate remains, and in 1906 he was in Egypt for the last time as a member of the Royal Prussian Academy of Science expedition for the study of the Nile.

Blanckenhorn represents the last of a generation of geologists whose devotion to work has made possible our knowledge of the geology of areas of the earth where travelling was a difficult, dangerous and annoying task. It needed giant and adventurous figures to cross unmapped and unknown regions of the inhospitable and barren Sahara by primitive means of travel and inadequate, if any, instruments for contacting the outside world.

The period of Blanckenhorn's travels was one of his most prolific and fertile periods. He published more than forty papers on this region. Blanckenhorn's ideas on the tectonics, structure, stratigraphy and paleontology of the Tertiary and Quaternary are still valuable. Many of his terms have leaked into the geological literature of that area. He later condensed his ideas in two books which appeared in the series "Handbuch der regionalen Geologie," namely, "Syrien, Arabien und Mesopotamien" (1914), and "Aegypten" (1921), published in Heidelberg. These two books still remain among the best written books on that area. At the end of this note a selected bibliography of Blanckenhorn's most important works is given.

Blanckenhorn became later an expert and co-worker of the "Preussischen Landesanstalt" where he examined and published the maps of the region between Kassel and Vogelsberg, and his experience has amplified our knowledge of the vulcanicity and tectonics of that region.

Blanckenhorn suffered during the two world wars. In the last war he lost most of his large library and was impoverished in the post war period. The present writer, an Egyptian geologist, was fortunate enough to be able to acquire the remains of his library and his personal geological instruments which will ultimately find their way to the Geological Museum of the Fouad I University, Cairo, Egypt.

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REVIEWS

The Chemistry and Technology of Enzymes; by HENRY TAUBER. Pp. viii, 550. New York, 1949 (John Wiley and Sons, Inc., $7.50). This book has been written to replace the Author's "Enzyme Technology." It combines in one volume both the industrial and the theoretical aspects of enzymes; thus meeting a need suggested by reviewers of the earlier volume. The first Part of the volume deals with the chemistry of enzymes in considerable detail (15 Chapters, 260 pages) with discussions of preparation, quantitative determination, and properties of them. The second part, on technology (17 Chapters, 259 pages) contains what appears to be a most up-to-date treatment of the industrial use of enzymes in the fermentation, baking and other food industries, medicinal, textile, leather, and other industries. The book reads well, has clear illustrations and appears to be thoroughly documented. The type and format give the book a good appearance. HAROLD G. CASSIDY

Dynamic Aspects of Biochemistry, American Edition; by ERNEST BALDWIN. Pp. xviii, 457. New York, 1947 (The Macmillan Company, $4.00. As the Author points out, biochemistry of necessity started from investigation of the chemical structure and configurations of the organic building blocks of living matter This might be called the morphological aspect of the subject, and is being continually advanced. Built upon this foundation, and of increasing importance and sublety are the investigations of the functions of these building blocks. This is the dynamical aspect of biochemistry. The volume is divided into two Parts: I, Enzymes, and II, Metabolism. The first two chapters of Part I give an excellent summary of the general aspects of enzyme behavior; the remaining three deal with specific groups of enzymes. The second Part contains a thoroughgoing discussion of the methods used in investigating intermediary metabolism; of the metabolism of carbohydrates, and of fats. There is a remarkable index of 37 pages in double columns. This book is designed as an advanced text. It should also be most valuable to one who wishes to read in the field. Not only will it refresh his memory and bring him up to date (1946) but it will provide the stimulating and idea provoking experience which those who have enjoyed the Author's "Comparative Biochemistry" (AM. JOUR. SCI. **237**, 365 (1939)) will expect. HAROLD G. CASSIDY

Annual Report on the Progress of Chemistry for 1947 (*Volume* 44); Issued by the Chemical Society. Pp. 327. London 1948 (The Chemical Society, 1). This volume, written by twenty experts summarizes the important advances which have been made in pure chemistry and related subjects up to the end of 1947 This, like its predecessors, is an extremely valuable volume because of its completeness and the general clarity and style which adorn it. The twenty topics are classified under the headings of General and Physical Chemistry; Inorganic Chemistry; Organic Chemistry; Biochemistry; Analytical Chemistry. The reader can thus gain, in a pleasant manner, a comprehensive view of many fields. For example, those interested in the earth sciences might find "The Kinetics of Electrode Processes" by J .N. Agar, which opens the volume, of equal interest with H. J. Dothie's "X-Ray Analysis" which is the penultimate topic. HAROLD G CASSIDY

Microwaves and Radar Electronics; by ERNEST C. POLLARD and JULIAN M. STURTEVANT. New York, 1948. Pp. vii, 426, 13.11 figs. (John Wiley and Sons, Inc., $5.00).—This book is a survey of microwaves and radar electronics, written for newcomers to the field who want to discover the potentialities of these new techniques. It should be particularly useful to two specific audiences—those who want to know what the microwave field is all about but who do not intend to use this particular volume as a technical manual, and those who seek a fluent familiarity with the nomenclature but can allow some one else to work out practical details.

The coverage of the book is good. The authors start with two chapters on fundamentals. They require an elementary knowledge of differential and integral calculus and vector notation, but the understanding of these two chapters is not prerequisite to an appreciation of later material. The third and fourth chapters under the titles of "The Production of Microwaves," and "Microwave Technique" give a general discussion of specifically microwave electronics. Chapters 5 to 11, comprising almost half the book, describe the electronic control circuits which have been designed for use with radar systems. These chapters should prove very useful to everyone interested in a summary of the chief advances made in electronic circuits during the war, particularly at the Radiation Laboratory at M.I.T. The description is sufficiently general so that the reader may easily recognize the value of these circuits in applications beyond the radar systems for which they were designed. Two chapters on radar and its accessories and microwave communications give brief explanations of the operational details of representative examples of these subjects.

The last chapter on "Microwaves in Physical Research" was somewhat of a disappointment. A consideration of the kinds of physical processes which might be investigated by the microwave method would have been more valuable than the discussion of specific examples to which the authors have limited themselves. It is obvious in a book on radar techniques that the emphasis should be on pulsed microwaves. The microwave techniques discussed, however, are not restricted to transient phenomena and particularly in physical research, the value of c-w magnetrons and other steady power sources is great It is unfortunate that this chapter on Physical Research does not point out clearly the possibilities of the steady source techniques as differentiated from the radar electronics.

The use of Radiation Laboratory slang makes the book unnecessarily difficult to read and robs it of considerable dignity. The inclusion of such terms as TR "boxes" for TR switch tubes, "the main bang" for trigger pulses, and "strips" for i-f amplifiers adds nothing to the clarity of presentation. Even an expert in the field may find confusing a quotation such as follows, "FTC and DBB circuits are most effective when used in conjunction with IAGC. The FTC and IAGC combination is effective for jamming by unmodulated c-w or c-w modulated at frequencies below the FTC cut off frequency, while the DBB and IAGC combination is best in cases of noise-modulated jamming and most types of clutter, especially clouds."

This reviewer believes that survey books make their real contribution by serving both as an introduction to the field and as a point of departure for the reader who needs specific and detailed information for a particular problem. Generally speaking, this book fails in the second objective, since many of the references are too indefinite to be useful. Too many of the footnotes read: "A detailed treatment will be found in the 'Radiation Laboratory Technical Series' published by McGraw-Hill Book Co.". Since there are twenty-seven volumes of this series, such a reference clarifies nothing. Although not all the volumes have yet appeared, the indexes of the unpublished volumes have been available for some years and a little investigation by the authors would have provided much more specific information. About half the references are to Radiation Laboratory Reports, and these are available only to a limited audience. Furthermore, as of January 1949, a check shows that 15% of the cited reports are still classified. It is clear that too little emphasis was placed on a proper selection of references to make them really useful.

SANBORN C. BROWN

Submarine Geology; by FRANCIS P. SHEPARD. Pp. xvi, 348; Bathymetric chart and 106 figs., New York, 1948 (Harper and Brothers, $6.00).—This addition to Harper's Geoscience Series by one of the most productive students of a new branch of geology should stimulate all students of earth science. Professor Shepard has made a valiant effort to dovetail the many related facets of oceanography and geology within a small volume. So rapidly are new techniques being devised and new data gathered, however, that the book is destined to become somewhat obsolete in the very near future, through no fault of the author, by virtue of geophysical advances. He clearly envisages this in a brief mention of equipment for the investigation of submarine geology in the Summary. The book is nonetheless a valuable contribution for all global-minded geologists pointing up as it does the unsolved problems of worldwide significance, solution of which depends on the methods (some yet to be devised) of submarine geology. The publishers have fashioned an attractive volume of high quality and pleasing format. The price seems high inasmuch as only one-third of the figures are halftones, most of them of small size.

The treatment is more that of a monograph than of a textbook, but this is probably inevitable in a field of inquiry so young and fastgrowing that a "standard" work will not be possible for many years. Oceanography was pursued very largely by non-American workers virtually up to World War II. Although he mentions a few such contributions in Chapter I, Shepard leans almost exclusively upon recent American work and freely admits (p. 9) embarrassment at the concentration of references to New England and California, areas in which he has specialized.

As Dr. Carey Croneis remarks in the Editor's Introduction, "some portions of the book may well come to be regarded as iconoclastic." This is due no doubt to Professor Shepard's keen enthusiasm and exuberance. The serious students will be stimulated but not misled by his Now It Can Be Told! style. Topic headings like "Some debunking" and "More debunking" in the discussion of continental shelves illustrate this. The author's enthusiasm occasionally carries him to sweeping generalizations. For example, under Significance of Glaciated Shelves one reads (p. 327) "Virtually all glaciated coasts have a superabundance of deep harbors, whereas harbors have to be constructed, or at least deepened, along most unglaciated coasts." What about, say, Rabaul, Tokyo, Manila, Rio and San Francisco, to say nothing of the great atolls of the Pacific such as Ulithi? In studies of continental shelves, particularly with submarine canyons, the author's experience is unequalled and the book reflects many original contributions.

Some conclusions, however, purporting to reflect results of recent studies by the author, are in fact so old that their origin is obscure. For example, in his summary of Waves, Currents and Engineering Structures (pp. 322-4) Shepard states: "The investigation of waves . . combined with recent studies of shore processes, particularly along the California coast, may provide a new basis for shore planning . . . . " He then lists sixteen "new discoveries, partly confirmation of principles not well understood previously" the majority of which have been known and utilized along heavily populated beaches of our east coast and in England for generations. The U. S. Beach Erosion Board has operated on these principles for nearly twenty years and before that state organizations like the New Jersey Board of Commerce and Navigation and State Geological Survey turned out detailed reports confirming these well-known principles.

A few errors have crept in, perhaps unavoidable in a new book. The already famous U. S. Hydrographic Office Chart No. 5485, first of a new series, is misnumbered (p. 3) 5495. Figure 96 showing seamounts (guyots) in the western Pacific is credited to H. O. Chart No. 2603 whereas it is actually a copy of Figure 1 in a paper by Hess published in this JOURNAL (Vol. 244, p. 774, 1946). Figure 33, a diagram of beach terminology, is uncredited although it appears identical with Figure 1 of the Interim Report of the U. S. Beach Erosion Board of 15 April 1933

SUBMARINE GEOLOGY is a welcome addition to geological literature. It briefly poses some of the greatest unsolved problems of geology, such as the origin of continental shelves, submarine canyons, flattopped seamounts and the Mid-Atlantic Ridge, conclusions about which are either scanty or highly controversial.

Chapter bibliographies invite the serious student to original sources of varying opinions. It should be of wide interest to all who know and love the sea and its shores, particularly to geologists and geophysicists engaged in its study and exploration. JOHN B. LUCKE

Foraminifera. Their Classification and Economic Use, with an Illustrated Key to the Genera; by JOSEPH A. CUSHMAN. Fourth edition, revised and enlarged. Pp. 605, 9 figs., 55 pls., Cambridge, Massachusetts, 1948 (Harvard University Press, $10.00).—The appearance of the fourth edition of this outstanding work marks its twentieth year of invaluable service to students of Foraminifera. The first ten chapters of the text, covering the general aspects of the subject, are preserved without appreciable change from the third edition. The remaining systematic portion of the volume is enlarged by the addition of analyses of 129 genera described since the third edition. The geologic range of many of the older genera has been changed in the light of recent work.

In revising the section on fusulines, Carl O. Dunbar has separated the group into two distinct families, the Fusulinidae and the Neoschwagerinidae, thus increasing the total number of families from 49 to 50. The section on the orbitoids by T. Wayland Vaughan and J Storrs Cole has been revised by Cole.

Seven new plates are added to the key to accommodate the illustrations of the additional genera. These illustrations maintain the high standards set in preceding editions. The bibliography has grown appreciably and includes a new section on Foraminifera as related to oceanography.

In continuing the use of the name *Camerina* and the family Camerinidae, Cushman seems to have overlooked the fact that under suspension of the *Rules*, the International Commission on Zoological Nomenclature in 1945 (Opinion 192) validated the name *Nummulites* and placed it on the official list of *nomina conservenda*. It should therefore replace *Camerina*.

Foraminifera has long been a standard paleontological reference of great popularity both here and abroad. A volume so well known and so widely acclaimed needs no further praise Dr. Cushman deserves the thanks of all students of Foraminifera for his efforts in keeping this excellent work up-to-date. KARL M. WAAGE

Bibliography of the Literature on Sodium and Iodine in relation to Plant and Animal Nutrition. Pp. 123. New York, 1948 (Chilean Nitrate Educational Bureau). This compilation of abstracts, primarily from Chemical Abstracts, on the literature of sodium and iodine in relation to nutrition is part of the more extensive collection of abstracts on the minor elements. It is a useful guide to the literature on the occurrence and physiological action of sodium and iodine. The index, however, is misleading since some elements are mentioned even if they merely are used as reagents in some analytical procedure. EVA M. LOW

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An Introduction to Chemical Science; by W. H. Hatcher. 2d Edition. New York, 1949 (John Wiley & Sons, $4.00).

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Biochemical Preparations. Vol 1. Herbert E. Carter, Editor-in-Chief. New York, 1949 (John Wiley & Sons, $2.50).

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THE BASE OF THE CAMBRIAN IN THE SOUTHERN APPALACHIANS

PHILIP B. KING

PART 1

ABSTRACT. The Cambrian differs from all other systems that succeed it in that its base is undefined paleontologically, for the Cambrian is underlain by stratified rocks without fossils. Recent publications have indicated considerable diversity of opinion as to the proper position of the base of the system.

As a contribution to the solution of the problem, the writer in this paper examines the stratigraphy of beds that have been assigned to the pre-Cambrian and early Cambrian in a typical area—the southern Appalachians. The oldest Lower Cambrian fossils in the area occur in the upper part of the Chilhowee group This group is a sequence of clastic rocks 2,500 to 7,500 feet thick, having closely related characters from base to top, and probably formed during a single depositional cycle It is interpreted as the initial deposit of the miogeocyncline in which the Cambrian and Ordovician rocks now exposed in the Appalachian Valley were deposited. Underlying the Chilhowee group locally are the Catoctin greenstone and its associated sedimentary rocks, the volcanics of the Mount Rogers area, and the Ocoee series. These rock units were formed under conditions of considerable crustal instability. and differ greatly in character from the formations of the Chilhowee group The Catoctin greenstone and the volcanics of the Mount Rogers area are separated from the Chilhowee group by erosional unconformities; an unconformity may also exist between the Ocoee series and the Chilhowee group All three of the rock units below the Chilhowee group rest with profound unconformity on ancient plutonic rocks.

It is concluded that there is a reasonable physical basis for drawing the base of the Cambrian in the southern Appalachians at the base of the Chilhowee group, even though the oldest Cambrian fossils appear well above this horizon The unconformity at the base of the Catoctin greenstone, the volcanics of the Mount Rogers area, and the Ocoee series—does not mark the base of the Cambrian but lies within the pre-Cambrian, serving to divide the pre-Cambrian rocks of the area into two parts of considerably different age.

[1] Published by permission of the Director, Geological Survey, U S Department of the Interior.

[2] While the present manuscript was passing through the hands of the Geological Survey editors, the following publication appeared: Stose, G W., and Stose, A. J, 1949 Ocoee series of the southern Appalachians. Bull Geol Soc Am, 60, 267-320 After reviewing this publication the writer concludes that it contains nothing which requires specific comment at this time, or necessitates modification of the conclusions presented in the present paper

STATEMENT OF PROBLEM

THE Cambrian differs from all other systems that succeed it in that its base is undefined paleontologically. At the boundaries of all succeeding systems, the faunas of one give place to those of the next, and however much the precise boundary may be debated, it is debated with knowledge of the faunas of the beds involved. In the Cambrian system, the lowest fossil zone is underlain by strata without fossils, or with fossils of so ambiguous a nature as to be useless for precise paleontological zonation.

Years ago, referring to the base of the Cambrian, Walcott (1891, p. 389) asked:

> "Is the basal line of the Cambrian group at the lowest limit at which the *Olenellus* fauna is found? Where this lowest limit occurs at the base of a conformable series resting unconformably upon pre-Cambrian rocks there is no difficulty in answering the question. But where it occurs in the midst of a conformable series and there remain thousands of feet of sediments beneath the *Olenellus* zone, as in Nevada and Utah, it still remains a problem for consideration."

The position of the base of the Cambrian system is not in dispute in widespread continental areas where Cambrian strata younger than the *Olenellus* zone lie on the eroded surface of much older igneous or metamorphic rocks of obvious pre-Cambrian age. But in geosynclinal areas such as those in the Appalachian and Cordilleran provinces, the fossiliferous Lower Cambrian is underlain in many places by unfossiliferous sedimentary or volcanic rocks. In the geosynclines, the questions posed by Walcott remain in dispute to this day.

Recent publications reflect the current state of uncertainty on the problem, and revisions of classifications commonly employed have been proposed by various geologists. However, these geologists are not in agreement among themselves, for some wish to place the base of the Cambrian higher, and others lower than the boundary commonly employed.

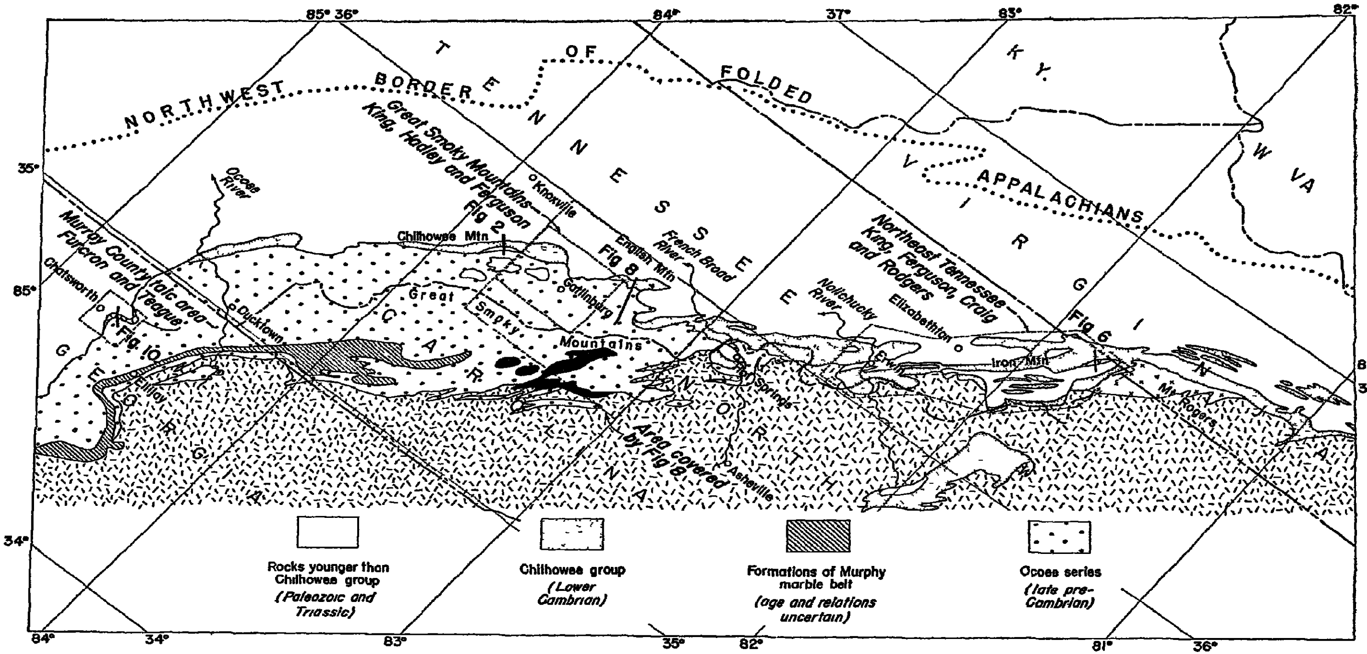
The Cambrian correlation chart, prepared by a subcommittee of the Committee on Stratigraphy of the National Research Council (Howell and others, 1944), shows at the bottom the lowest formation that contains diagnostic Cambrian fossils, but does not show other formations beneath, even though these are also conventionally classed as Cambrian. Snyder (1947, p. 152) and Wheeler (1947, pp. 157-159) express approval of this procedure, and recommend that the base of the Cambrian be placed either at the base of the lowest formation containing Cambrian fossils, or immediately beneath the lowest occurrence of Cambrian fossils.

By contrast, Bloomer and Bloomer (1947, pp. 102-106) conclude that the volcanics and underlying sedimentary rocks in Virginia are of Cambrian age, even though they lie some distance beneath fossiliferous Cambrian strata and have been classed conventionally as pre-Cambrian. This conclusion is based on a lithologic similarity between beds hitherto classed as Cambrian and as pre-Cambrian, and on a lack of certain unconformity between them. Likewise, Furcron and Teague (1947, pp. 19-21) conclude that the unfossiliferous Ocoee series of northern Georgia, hitherto variously classed as Cambrian or pre-Cambrian, is actually Cambrian because it rests unconformably on older granites and gneisses.

PURPOSE OF THIS PAPER

The southern Appalachians is one of the critical areas in which the problem of the position of the base of the Cambrian comes to the fore, and conclusions drawn on the basis of study of this area have a bearing on the general solution. In the present paper, the writer proposes to examine the stratigraphy of those beds in this region that have been classed as pre-Cambrian and early Cambrian, and to offer some tentative suggestions as to their classification. The paper is an outgrowth of work in progress and ideas in process of evolution. It states a thesis which will be elaborated and perhaps modified as this work proceeds.

During the past eight years (1940-1948) the writer has participated in three field projects of the U. S. Geological Survey, involving study of older rocks of the southern Appalachians: (1) Near Elkton, in Page and Rockingham Counties, northern Virginia, for a distance of 20 miles along the strike of the rocks in the northwestern foothills of the Blue Ridge (King, 1943, 1947) (fig. 1). (2) In Johnson, Carter, and Unicoi Counties, northeastern Tennessee, for a distance of 40 miles along the strike of the rocks in Iron, Holston, and Unaka Mountains (King and others, 1944). (3) Near Gatlinburg, in Blount and Sevier Counties, east-central Tennessee,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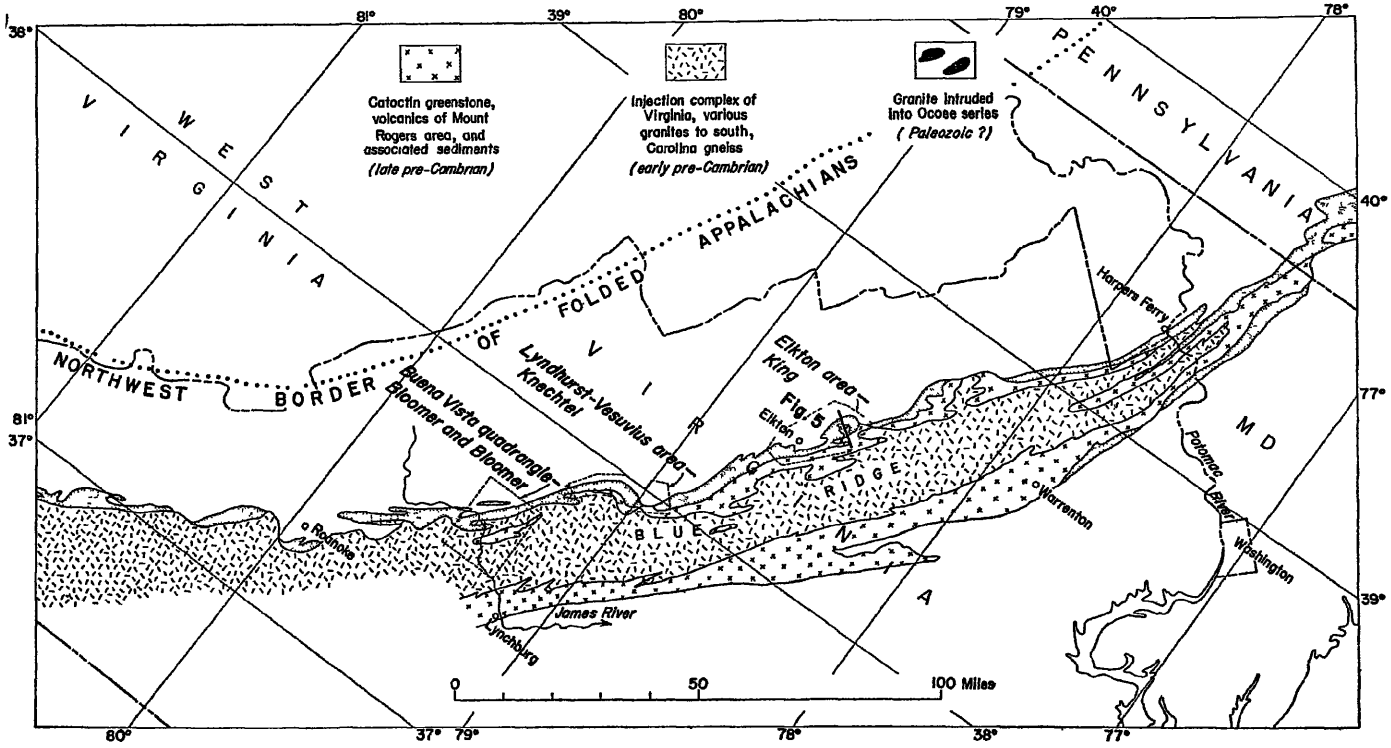
NORTHWEST BORDER OF FOLDED APPALACHIANS
TENNESSEE
KY.
W VA
VIRGINIA
NORTH CAROLINA
GEORGIA
Great Smoky Mountains—King, Hadley and Ferguson
Fig 2
Northeast Tennessee—King, Ferguson, Craig and Rodgers
Fig 6
Murray County talc area—Fulcron and Teague
Fig 10
Area covered by Fig 8
Fig 8
Ocoee River
Knoxville
Chilhowee Mtn
English Mtn
French Broad River
Nolichucky River
Elizabethton
Iron Mtn
Mt Rogers
Gatlinburg
Great Smoky Mountains
Ducktown
Chatsworth
Ellijay
Springs
Asheville
Rocks younger than Chilhowee group (Paleozoic and Triassic)
Chilhowee group (Lower Cambrian)
Formations of Murphy marble belt (age and relations uncertain)
Ocoee series (late pre-Cambrian)

Figure 1. Sketch map of parts of Virginia, Tennessee, and North Carolina, showing outcrops of units of early Cambrian and late pre-Cambrian age, and location of places mentioned in text Areas enclosed in rectangles have been covered in recent geologic investigations that are referred to in text.


Catoctin greenstone, volcanics of Mount Rogers area, and associated sediments (late pre-Cambrian)
Injection complex of Virginia, various granites to south, Carolina gneiss (early pre-Cambrian)
Granite intruded into Ocoee series (Paleozoic ?)
WEST VIRGINIA
PENNSYLVANIA
MD
VIRGINIA
NORTHWEST BORDER OF FOLDED APPALACHIANS
BLUE RIDGE
Elkton area—King
Fig. 5
Elkton
Lyndhurst-Vesuvius area—Knechtel
Buena Vista quadrangle—Bloomer and Bloomer
Harpers Ferry
Warrenton
Potomac River
Washington
Roanoke
Lynchburg
James River
0
50
100 Miles
38°
81°
39°
80°
79°
40°
78°
77°
37°

for a distance of 22 miles along the strike of the rocks in the Great Smoky Mountains and Chilhowee Mountain. This last investigation is now in progress. These areas are widely scattered, but each is a representative sample of the broader district of which it forms a part. The writer's knowledge of intervening areas is based on field excursions and on a review of published papers.

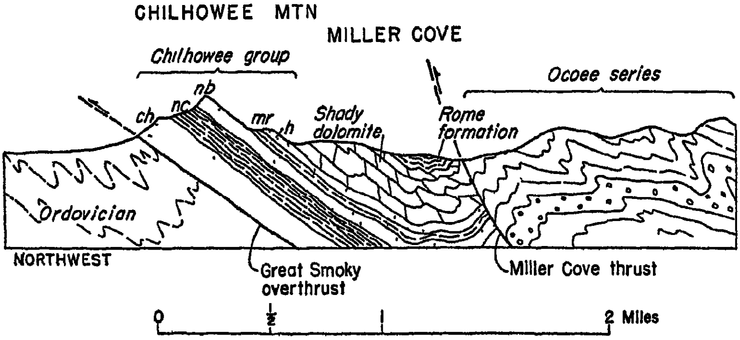


Figure 2. Structure section of Chilhowee Mountain and Miller Cove, Blount County, Tennessee, showing type section of Chilhowee group. Formations of Chilhowee group are indicated by abbreviations as follows: ch—Cochran conglomerate, nc—Nichols shale, nb—Nebo quartzite, mr—Murray shale, h—Hesse quartzite. After Keith, with modifications by the writer

During the investigations in the southern Appalachians the writer has been associated, among others, with John Rodgers of Yale University, H. W. Ferguson of the Tennessee Division of Geology, and J. B. Hadley of the U. S. Geological Survey. Some of the field facts given in this paper were determined by them rather than by the writer, and others were determined jointly in their company in the field. Interpretations and conclusions given here have been discussed with them, both before and during preparation of this paper, but the theses presented herein are the writer's own, and are not necessarily shared by his colleagues.

CHILHOWEE GROUP

Terminology.—One of the most widespread units of the southern Appalachians is the Chilhowee group, named by Safford (1856, pp. 152-153; 1869, pp. 198-203) for Chilhowee Mountain, Blount and Sevier Counties, east-central Tennessee (fig. 2).[3] The group underlies the Cambrian and Ordovician carbonate rocks (Shenandoah limestone of early reports) of the Appalachian geosyncline, which form the Appalachian Valley. The Chilhowee group consists of clastic rocks 2,500 to 7,500 feet thick that stand as ridges and mountains southeast of the valley (fig. 1). Representatives of the group occur in Tennessee and Virginia, and equivalents extend southwest into Alabama and northeast beyond Pennsylvania. The Chilhowee is conventionally classed as the lowest unit of the Lower Cambrian series.

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[3] Problems relating to usage of the term Chilhowee group are discussed under the heading "Notes on terminology" at the end of this paper.

The Chilhowee group has been subdivided into different sets of formations in its several areas of exposure; these variations in terminology obscure the essential unity of the group, the wide extent of its characteristic rock types, and the similarity of sequence in all sections. The terminology in three principal areas of outcrop is indicated in table 1.

TABLE 1

Formations of the Chilhowee group in Tennessee and Virginia[4]

		East-central Tennessee (Chilhowee Mountain)	Northeastern Tennessee (Johnson, Carter and Unicoi Counties)	Northern Virginia (Elkton and Harpers Ferry areas)
Lower Cambrian		Shady dolomite (in Miller Cove)	Shady dolomite	Tomstown dolomite
	Chilhowee group	Hesse quartzite	Erwin quartzite	Antietam quartzite
		Murray shale		
		Nebo quartzite		
		Nichols shale	Hampton shale	Harpers shale
		Cochran conglomerate	Unicoi formation (with basalt flows 1000–1500 feet below top)	Weverton quartzite
				Loudoun formation (with tuffaceous slate and rare flows)
Pre-Cambrian		Ocoee series	Volcanics of Mt. Rogers area Cranberry granite	Catoctin greenstone Swift Run formation Injection complex

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[4] See also "Notes on terminology" at the end of this paper.

Fossils.—Fossils are found sparingly in the upper part of the Chilhowee group. Small collections of trilobites and brachiopods have been made from the topmost beds, that is, from calcareous sandstones at the top of the Hesse, Erwin, and Antietam quartzites (Butts, 1940, p. 40). These fossils include *Olenellus* and are of Lower Cambrian age. In addition, Keith (1895, p. 3) reported the occurrence of trilobites and brachiopods at a horizon nearly 500 feet lower, at two localities ascribed to the Murray shale of Chilhowee Mountain: (1) On the east side of Little River Gap, and (2) on the crest of the mountain above Montvale Springs. The first of these stations is of doubtful stratigraphic position and may well be in the top beds of the Chilhowee group (Stose and Stose, 1944, p. 388), but the second is almost certainly in the Murray shale and represents the oldest recorded occurrence of trilobites and brachiopods in the southern Appalachians. Most of the fossils from the Chilhowee group were collected years ago by Walcott (1891, p. 302; 1892; 1892A) and further detailed search might yield other localities and horizons, as well as greater knowledge of these oldest Cambrian faunas.

Besides the trilobites and brachiopods, the Chilhowee group contains other indications of life—*Scolithus*, or probable worm tubes. These occur at approximately the same level as the trilobites and brachiopods, but also extend for a considerable distance beneath them. The tubes are exceedingly abundant in parts of the Antietam and Erwin quartzites (Stose and others, 1919, pp. 16-17; King and others, 1944, pp. 30-31). In the Erwin, they are especially abundant in white quartzite beds in the lower part of the formation. They are similarly abundant in the Nebo quartzite. *Scolithus* also occurs in indurated sandstone beds in the Hampton and Harpers shales. It is absent in the lower formations of the Chilhowee group.

Conditions of deposition.—The Chilhowee group is a conformable sequence of beds, formed during a single depositional cycle, with no great difference in age between its component parts. All sections possess a similar upward sequence of deposits. The most distinctive parts of the group, the coarse clastics, change from arkosic sandstones and arkosic conglomerates below (in the Cochran, Unicoi, and Weverton formations) to fine-grained, indurated, quartzose sandstones (generally termed "quartzites") above (in the Nebo, Hesse, Edwin, and Antietam formations).[5] The conglomerates of the lower beds contain rounded pebbles and granules of quartz and other resistant constituents of older rocks, but these are absent above. Cross bedding ("current bedding" of British geologists) is common, and ripple marks are not rare. Sediments of the Chilhowee group were probably derived from lands of moderate relief, which by the end of Chilhowee time had probably been peneplaned. Sediments of the Chilhowee group are the initial deposits of the miogeosyncline (Kay, 1947, p. 129) in which the Cambrian and Ordovician rocks now exposed in the Appalachian Valley were deposited.

The occurrence of fossils in the Chilhowee group is closely related to the sequence of sediments: the lower arkosic and conglomeratic part is barren; the middle part, with its quartz sand deposits contains only *Scolithus*; and the upper part, laid down in the waning stages of sand deposition contains trilobites and brachiopods. This relation so closely resembles the relation in later transgressive deposits as to suggest that the appearance of diagnostic fossils in the upper part of the Chilhowee group is due, not to any abrupt evolutionary development, but to migration of marine life into the region in response to an increasingly more favorable environment.

A distinctive feature of the Chilhowee group is the presence of volcanics which occur in some areas. In northeast Tennessee and southwest Virginia the Unicoi formation, the basal formation of the group, contains flows of amygdaloidal basalt and associated tuffaceous shales and sandstones (fig. 6) (King and others, 1944, pp. 39-40; Butts, 1940, pp. 27-31). Great variations in the volcanics of the Unicoi may be observed both along the strike and between different outcrop belts. In places, two or more volcanic beds are present, each 100 or more feet thick; in others a single thin bed is present; in others volcanic rocks are lacking. The volcanics in the Unicoi pinch out not far southwest of the Nolichucky River in Tennessee, and are absent from the Unicoi and its stratigraphic equivalent, the Cochran conglomerate, farther southwest. These variations arise from the expectable conditions of a volcanic episode.[6]

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[5] For explanation of rock names used, see "Notes on terminology" at end of this paper.

In northern Virginia at the base of the Chilhowee group are tuffaceous spotted slates and rare thin beds of lava (Furcron and Woodward, 1936). These volcanics are a part of the Loudoun formation (Stose and Stose, 1946, pp. 31-34); in the Elkton area they comprise the whole of the formation, but near the Potomac River they are interbedded with sandstone and arkose. In the Buena Vista quadrangle to the southwest, volcanics also form the base of the Chilhowee group (Bloomer and Bloomer, 1947, p. 103) and may be the same unit as the Loudoun, although they are there termed "the basal portion of the Unicoi."

Correlation within the Chilhowee group.—The top of the Chilhowee group or its contact with the overlying Shady and Tomstown dolomites, seems to be a well-defined time horizon of about the same age everywhere, and expresses a marked, widespread change in sedimentation (table 1). Correlation of the underlying parts of the group is less certain. The Unicoi of northeastern Tennessee usually has been correlated with the Loudoun and Weverton of northern Virginia, the Hampton with the Harpers, and the Erwin with the Antietam. All these units are lithologic, and are merely the local expression of upward changes in sedimentation during Chilhowee time, such as that from arkosic sandstone and conglomerate below to quartz sandstone above.

There is no assurance that these lithologic units were laid down at the same time in all areas. The Chilhowee group thins from a maximum of 7,500 feet in parts of Tennessee to 2,500 feet near the Potomac River in northern Virginia. Conceivably the Weverton quartzite, or northern representative of the arkosic sandstone facies, might have been deposited at a later time than the Unicoi, or southern representative of the same facies (fig. 3B).

The volcanics of the Chilhowee group afford some clue as to correlation within the group. The volcanics in the Unicoi formation of northeast Tennessee are quite variable from one local outcrop to another, but they culminate in a zone that is 1,000 to 1,500 feet below the top of the Unicoi, or 4,000 to 4,500 feet below the top of the Chilhowee (they are hereafter referred to as the "main volcanic zone" of the Unicoi; "main basalt" of fig. 6). Those in the Loudoun formation of northern Virginia are at the base of the Chilhowee group and 2,500 feet below its top. The writer believes that the volcanics of the Unicoi express an episode of regional time significance, and he suggests that those of the Loudoun formation are of approximately the same age. If this correlation is accepted, the Weverton, Harpers, and Antietam are probably the thinned equivalents respectively of the upper Unicoi, the Hampton, and the Erwin (fig. 3A), rather than a similar facies of later age (fig. 3B).

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[6] On these grounds, the writer disagrees with Stose and Stose (1944, p. 389) who state, "With the present knowledge, it is not possible to decide which formation, if any, in the Chilhowee type section represents the Unicoi formation, but *it is evident that the lower part of that formation which contains the basalt flows is not present* because neither Hayes nor Keith describe such flows in the section" (italics the present writer's).

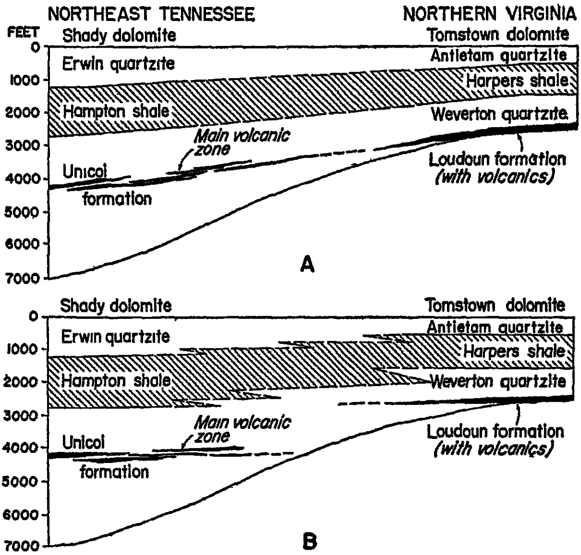

Figure 3 Stratigraphic diagrams between northeast Tennessee and northern Virginia, indicating possible correlations: (A) on the assumption that the volcanics in the two areas are of the same age, and (B) on the assumption that they are of different ages

By either of the interpretations just suggested, the sediments below the main volcanic zone of northeast Tennessee, classed as a lower part of the Unicoi formation, are older than any part of the Chilhowee group of northern Virginia, and are not represented there. As no unconformity is known between them and the upper part of the Unicoi, they seem to be a part of the Chilhowee depositional cycle. These lower beds may therefore pass out by overlap toward the northeast. Further comment on these beds is made below.

Correlation between formations of the Chilhowee group of northeast Tennessee and those of Chilhowee Mountain farther southwest will now be considered. No special problems seem to be involved in the beds above the Cochran conglomerate and Unicoi formation, and relations as indicated in table 1 may be regarded as reasonably well established. The Nebo quartzite of Chilhowee Mountain, for example, bears abundant *Scolithus* as does the apparently equivalent lower member of the Erwin quartzite of northeast Tennessee.[7] Also, H. W. Ferguson (personal communication, 1948) has found in the Nichols shale at a few places near the northeast end of Chilhowee Mountain the same characteristic arkosic sandstones that occur interbedded in the Hampton shale of northeast Tennessee.

More uncertainty exists as to correlation of the Unicoi formation and the Cochran conglomerate. In northeast Tennessee, Keith (1907A, p. 5 and areal geology sheet) placed the equivalent of the Cochran high in the section, at some localities in what is classed by King and others (1944) as upper Unicoi, in others as lower Hampton. Lower beds of the Unicoi were correlated with the Snowbird formation and Hiwassee slate, here regarded as units of the Ocoee series. The Cochran is an arkosic sandstone and conglomerate of the same facies as the upper part of the Unicoi. It is reported to be 1,100 to 1,700 feet thick on Chilhowee Mountain (Keith, 1895, p. 3), but exceeds 2,500 feet near the French Broad River to the east (Keith, 1904, p. 5; H. W. Ferguson, personal communication, 1948).

Sections of the Unicoi formation in northeast Tennessee show wide variations in thickness and considerable differences in lithology, especially within the strata below the main volcanic zone, classed as the lower part of the Unicoi—a variation emphasized by the piling up of thrust sheets, and the consequent juxtaposition of masses of Unicoi originally deposited far apart. Beds originally laid down farthest to the northwest are exposed along the Nolichucky River southeast of Erwin (fig 1) and along the Doe River south of Hampton 20 miles to the northeast. At these places the Unicoi is 2,000 to 2,500 feet thick and includes 1,000 feet or more of beds older than the main volcanic zone. These lower strata are arkosic sandstones very much like the upper strata, and the whole unit is probably equivalent to the Cochran. The possibility seems remote that any of it is equivalent to the Snowbird or other formations of the Ocoee series, as suggested by Keith (1907A).

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[7] In a recent paper Stose and Stose (1947, p. 629) interpret the Nebo as equivalent to the upper part of the Unicoi of northeast Tennessee, but there is little basis for this view.

Other occurrences of the Unicoi formation in northeast Tennessee, notably those of Iron Mountain (fig. 1), represent parts of the farther traveled thrust sheets, and show thicknesses of Unicoi of as much as 5,500 feet (fig. 6). On Iron Mountain, the beds below the main volcanic zone consist "largely of thin-bedded arkosic siltstone and sandstone, but layers of red and green clay shale and poorly indurated conglomerate are also present" (King and others, 1944, p. 40). These beds possess more of a graywacke facies than an arkosic sandstone facies and, as pointed out by Stafford (1869, pp. 195-196) resemble the rocks of the Ocoee series. At least the lower part of these beds is probably older than the type Cochran conglomerate, as well as of the Loudoun formation and Weverton quartzite to the northeast. Whether any of them is equivalent to the true Ocoee series farther southwest is problematical.

CATOCTIN GREENSTONE AND UNDERLYING SEDIMENTARY ROCKS

In northern Virginia, and in Maryland and Pennsylvania as well, the most prominent formation beneath the Chilhowee group is the Catoctin greenstone[8] (Keith, 1894, pp. 306-309), a series of thick, massive lava flows and associated agglomerates, tuffs, and tuffaceous slates. The Catoctin has generally been classed as a metabasalt (Furcron, 1934, pp. 401-404; Stose and Stose, 1946, pp. 20-22), but in the Buena Vista quadrangle Bloomer and Bloomer (1947, p. 97) have identified it as a propylite, or altered andesite. In northern Virginia, Maryland, and Pennsylvania, the greenstone is interbedded with flows of rhyolite and andesite (Keith, 1894, pp. 302-306; Stose and Stose, 1946, pp. 22-24).

The Catoctin greenstone is nearly everywhere underlain by sedimentary formations of variable thickness which include the Swift Run formation from Elkton north to the Potomac River (Stose and Stose, 1946, pp. 18-20; King, 1947), and its probable correlatives — the Fauquier formation of Furcron (1939, pp. 37-41) in the Warrenton area, and the Oronoco formation of Bloomer and Bloomer (1947, p. 95) in the Buena Vista area (fig. 1).

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[8] See "Notes on terminology" at end of this paper.

Bloomer and Bloomer (1947, p. 103) report that "without the aid of field relations it is difficult if not impossible to distinguish the rocks of the Oronoco formation from those in the lower part if the Unicoi formation." By contrast, the Swift Run formation of the Elkton area is greatly different from the younger Weverton quartzite. Both the Swift Run and Weverton include much arkosic sandstone, but the most prominent detrital constituents of the Swift Run formation are pink feldspar and epidote, derived from the injection complex beneath it and lacking in the Weverton. Quartz grains and pebbles, including much "blue" quartz, are more prominent in the Weverton quartzite than in the Swift Run formation. The two formations seem to have been laid down under different geographic conditions, and may have been derived from different sources.

The relation of the Catoctin greenstone and underlying sedimentary rocks to the plutonic rocks of the injection complex has been much disputed. Keith (1894, pp. 314, 318) interpreted the injection complex as intrusive into the Catoctin, and this view has been reaffirmed by Furcron (1934, pp. 405-410) and Cloos (Cloos and Hietanen, 1941, pp. 80, 83, fig. 29). However, Jonas and Stose (1939, pp. 579-580) and Bloomer and Bloomer (1947, pp. 95-97) consider that the Catoctin and the underlying sedimentary rocks lie unconformably on the injection complex. The latter view has been confirmed by the writer's own observations in the Elkton area (King, 1947). Many outcrops were found, of which the most accessible is on the south side of the East Branch of Naked Creek a mile east of Jollett (fig. 4), which display the rocks of the Swift Run formation lying unconformably on the eroded surface of the injection complex. The indurated sandstones of the Swift Run formation of the Elkton area contain minerals derived from the erosion of the injection complex. The Swift Run persists throughout the Elkton area as a nearly continuous stratum 50 or 100 feet thick between the Catoctin above and the injection complex below, an unlikely relation if the latter were intrusive into the former.

The Catoctin greenstone and underlying sedimentary rocks thin out in places, so that the overlying Chilhowee group overlaps them and lies directly on the injection complex. This relation was observed by Stose and Stose (1946, pp. 28-29) from Purcell Knob, northern Virginia, northward across the Potomac into Maryland. It was observed by the writer in the Shenandoah salient, in the northeast part of the Elkton area (fig. 5). It was observed farther southwest in Virginia by Knechtel (1943, pl. 29) in the Lyndhurst-Vesuvius area (fig. 1), and by Bloomer and Bloomer (1947, pp. 95-97) in the Buena Vista quadrangle. All the localities cited are along the northwest edge of the Blue Ridge; in the Blue Ridge itself the Catoctin is a thick and continuous body. The observed wedging out of the group is therefore probably not a set of unrelated local features, but represents a regional boundary

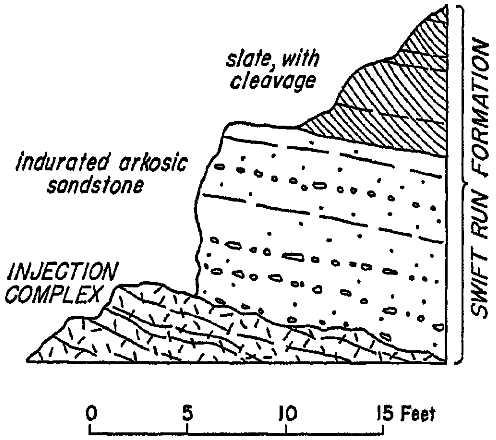

Figure 4 Sketch section of outcrop of Swift Run formation on south bank of East Branch of Naked Creek a mile east of Jollett, Page County, Virginia, showing relations with injection complex.

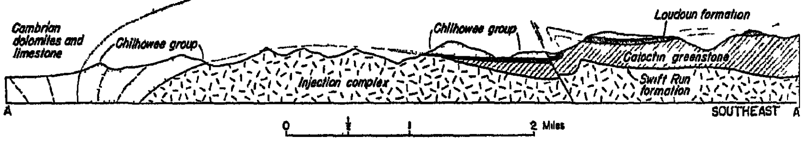

Figure 5 Geologic section across Shenandoah salient, Page County, Virginia, northeast of Elkton, showing relations of injection complex, Swift Run formation, and Catoctin greenstone to Chilhowee group. Based on observations by the writer.

along which the Chilhoweee extended northwest across the edge of the Catoctin.

Bloomer and Bloomer (1947, pp. 104-105) interpret the passage from Catoctin to Chilhowee time as marked by a shift in the position of the basins of deposition, but believe that the relations are essentially conformable. However, in the Buena Vista quadrangle, the Catoctin greenstone and Oronoco formation are not in contact with the Chilhowee group, and occur in a different fault block, so the critical wedge relations at their edge cannot be observed. In the Shenandoah salient of the Elkton area, the wedge relations can be observed, and the Catoctin and Swift Run are definitely unconformable beneath the Chilhowee (fig. 5). Within the space of a few miles the Catoctin thins abruptly northwestward, the underlying Swift Run formation approaches the Chilhowee, and where the Catoctin is absent, lies directly beneath it. Still farther northwest, both the Catoctin and Swift Run are missing and the Chilhowee lies on the injection complex.

Erosion of the edges of the Catoctin greenstone before and during Chilhowee time is indicated by the common occurrence in the Weverton quartzite of jasper pebbles derived from amygdules in the lavas of the Catoctin (Keith, 1894, pp. 312-313). If the correlations within the Chilhowee group suggested earlier in this paper are correct, the unconformity between the Catoctin and the Chilhowee in northern Virginia would involve a hiatus of at least sufficient length to permit the deposition in northeast Tennessee of that part of the Unicoi older than the main volcanic zone.

The writer agrees with Bloomer and Bloomer that the change from Catoctin to Chilhowee time involved a shift in the position of the basin of deposition, but he believes that this also involved tilting of the edges of the older basin, erosion of the marginal beds of the basin, and probably a considerable time hiatus.

VOLCANICS OF MOUNT ROGERS AREA

Southwest of the James River on the northwest flank of the Blue Ridge, the Catoctin greenstone and underlying sedimentary rocks are absent, and the Chilhowee group lies for a distance of 120 miles on rocks of the injection complex (fig. 1). If any intervening sedimentary or volcanic rocks once existed, they were removed by erosion prior to the deposition of the Chilhowee group.

In the vicinity of Mount Rogers, immediately north of the Virginia-Tennessee line, another body of volcanics lies beneath the Chilhowee group (Jonas and Stose, 1939, pp. 590-591). These volcanics are dominantly rhyolitic, whereas rhyolite is subordinate in the area of the Catoctin greenstone, but they are considered to be the approximate equivalent of the Catoctin by Jonas and Stose (1939, p. 575). The volcanics of the Mount Rogers area are unconformable beneath the Chilhowee group, for "north of White Top these [volcanic] conglomerates dip under, and are parallel to, the Lower Cambrian, Unicoi formation, but at the only contact observed there was a slight angular unconformity between the two series. Eastward the beds of the volcanic series trend at an angle with the strike of the Unicoi formation and still farther east the volcanic series is not present and the Unicoi formation rests on the injection complex" (Jonas and Stose, 1939, p. 591).

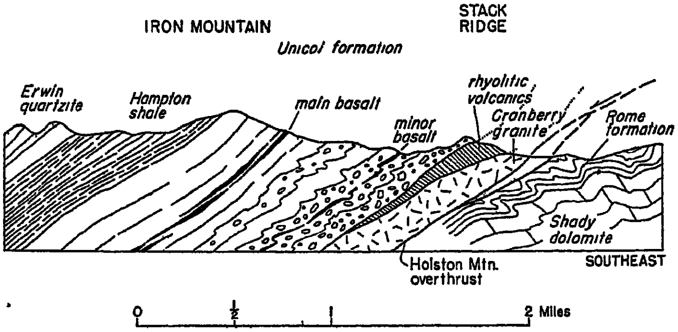

Figure 6 Geologic section through Iron Mountain and Stack Ridge, Johnson County, Tennessee, showing relations of Unicoi formation to rhyolitic volcanics and to Cranberry granite The rhyolitic volcanics are a part of the volcanics of the Mount Rogers area Based on observations of the writer and of H. W. Ferguson of the Tennessee Division of Geology.

The southwest edge of the volcanics of the Mount Rogers area extends across the Virginia-Tennessee line into Johnson County, northeast Tennessee, where relations similar to those just described have been observed by the writer and his colleagues (King and others, 1944, pp. 41-42, pl. 3). H. W. Ferguson has mapped rhyolitic lavas and indurated tuffs in the southeastern foothills of Iron Mountain—on Fodderstack Mountain and Stack Ridge (fig. 6). The volcanics are overlain by conglomerates of the Unicoi formation which contain pebbles derived from the volcanics. On Stack Ridge the volcanics are a southwest tapering wedge between Cranberry granite below and Unicoi formation above. The wedge terminates at the southwest end of the ridge, and farther southwest in Tennessee the Unicoi lies directly on the Cranberry and other granites.

In the Mount Rogers area, as in northern Virginia, the volcanics are bodies of variable thickness lying beneath the Chilhowee group, which were apparently truncated by erosion before the beginning of Chilhowee time. As nearly 3,000 feet of the lower beds of the Unicoi intervene between the volcanics of Stack Ridge and the main volcanic zone of the Unicoi exposed on the slopes of Iron Mountain, it is evident that the two volcanic episodes are separated by a considerable time interval.

(*To Be Continued*)

THE FAULT AT THE WEST EDGE OF THE TRIASSIC IN SOUTHERN PENNSYLVANIA

GEORGE W. STOSE

ABSTRACT The floor of the Triassic basin in southern Pennsylvania is exposed in several places close to the western margin of the basin where the thickest sediments would be expected. To explain this anomaly, a fault is postulated within the basin at the west edge of the deeper part A line of stocks, representing magma that rose from a deep source, is believed to mark the line of the postulated fault.

IN southern Pennsylvania a longitudinal series of normal faults bounds the west edge of the Triassic, and is here referred to as the Triassic border fault. The Triassic sedimentary rocks, comprising arkosic sediments below—New Oxford formation—and the Gettysburg shale above, dip uniformly northwest at an average rate of about 30°. The Triassic formations were deposited in a basin which is believed (Stose, 1932, fig. 5, p. 54) to have progressively deepened westward, due to gradual sinking of the basin on a fault at the west margin, and the sedimentary layers overlapped each other westward. These Triassic formations have a combined aggregate thickness of about 23,000 feet in this area (Stose, 1929, pp. 8-10), and the greatest accumulation of sediments should occur at or near the west edge of the basin. Here the youngest beds of the Triassic are quartzose and limestone conglomerates derived from the waste of uplifted rocks northwest of the border fault (fig. 1). The presence of Paleozoic limestone forming the floor beneath these youngest Triassic rocks at and near the west edge of the basin in the Carlisle quadrangle (fig. 2), therefore, is not to be expected and requires explanation.

In the Carlisle quadrangle along the normal faults at the west edge of the Triassic, the youngest Triassic sedimentary beds in the down-dropped block comprise quartzose and limestone conglomerates, and the uplifted rocks to the west consist largely of late pre-Cambrian volcanic rocks. The Triassic conglomerates were formed from the waste of rocks then present west of the fault in the closing stages of Triassic deposition. The presence of quartzose conglomerate east of the fault indicates that the volcanic rocks northwest of the fault were still covered by Cambrian quartzite when the conglomerate accumulated. These quartzose conglomerates pass northward into limestone conglomerate in the vicinity of Dillsburg, hence at the time of formation of this conglomerate Cambrian limestone capped the quartzite west of the bounding fault.

The limestone that forms the floor of the Triassic sediments east of the border fault (fig. 2) crops out in the valleys of South and West branches of Latimore Creek northwest of York Springs, and underlies the Triassic sedimentary rocks and an associated diabase flow or sill for a distance of 3 miles across the strike to the normal fault at the west border. The contact between the limestone floor and the overlying Triassic beds is approximately horizontal. Another area of the limestone floor beneath nearly horizontal Triassic limestone conglomerate and red sandstone in the Carlisle quadrangle is near the border fault in the lowland west of Dillsburg. A similar area of Paleozoic limestone east of the border fault forms the floor of Triassic conglomerate and sandstone just north of

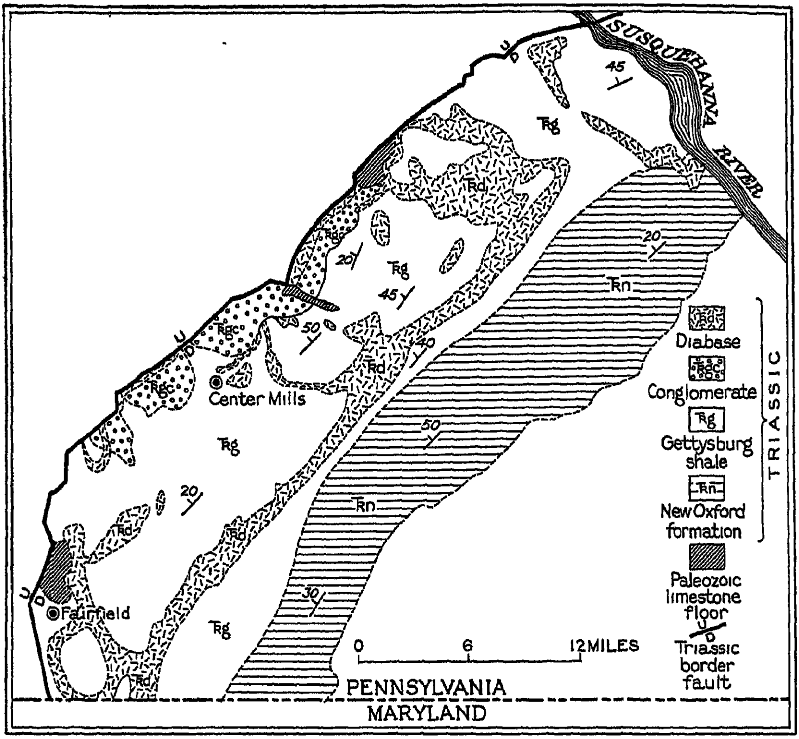

Fig 1. Map of the Triassic rocks in southern Pennsylvania.

Fairfield (fig. 1). These exposures of the limestone floor were at first explained (Stose and Lewis, 1916, p. 629) as due to the overlapping of Triassic sedimentary beds in a progressively westward sinking basin, but this interpretation was abandoned after detailed mapping of the area (Stose, 1929, p. 16).

Diabase intrusions in the Triassic sedimentary rocks consist of dikes, sills, large cross-cutting bodies, and stocks.

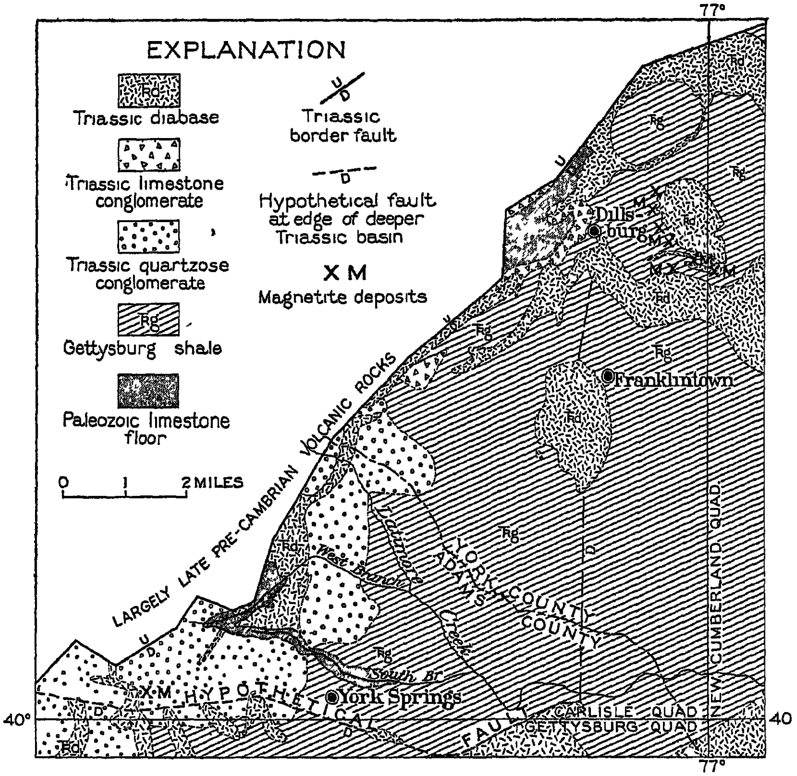

Fig. 2 Geologic map of part of the Carlisle quadrangle, Pa, and adjacent area

Those who have studied these intrusive rocks and the associated magnetite deposits (Spencer, 1908, pp. 76-96; Harder, 1910, pp. 612-618; Smith, 1931, pp. 32-52) agree that the diabase originated in a deep-seated magmatic reservoir below the Triassic rocks, ascended through these rocks along vents, and are represented by stocks and large cross-cutting bodies. The large cross-cutting bodies (fig. 1) do not extend to the west limit of the Triassic rocks but end east of the exposures of the limestone floor at Fairfield, York Springs, and Dillsburg. A large cross-cutting body east of the limestone floor exposed at Fairfield ends about 1 mile east of the border fault. The diabase bodies south of York Springs, along the south margin of the Carlisle quadrangle and the northern part of the Gettysburg quadrangle extending south to Central Mills, are cross-cutting bodies and are believed to represent a partially exposed large stock that lies south of the limestone floor at York Springs. The oval-shaped area of diabase at Franklintown appears to be a stock, and the large cross-cutting bodies southeast and east of Dillsburg are stocks that lie east of the limestone floor exposed in the vicinity of Dillsburg.

These cross-cutting bodies of diabase have a large content of original magnetite, which was a component of the deep-seated magma, and the occurrence of concentrated magnetite deposits in the metamorphosed rocks adjacent to most of these stocks, and deposits of garnet in the metamorphosed rocks, therefore, are believed to indicate that these stocks had direct connection with the deep source of the molten magma. It is suggested that the magma ascended from the deep source along faults in the floor of the Triassic basin at the west edge of a deeper part of the basin. Such a fault would extend to the bottom of the thick sediments accumulated in the basin, at a depth of about 20,000 feet, and deeper into the rocks of the floor beneath, and would provide a direct channel up which the magma could rise as it was squeezed out from beneath the sinking basis. This fault, which marks the west edge of the deeper basin, would be located at a line joining the northwest ends of the cross-cutting bodies and east of the limestone floor exposed at Dillsburg, York Springs, and Fairfield (fig. 2). Further evidence of the existence of such a fault is the fact that the Triassic beds to the east dip uniformly northwest, whereas those to the west are practically horizontal. The suggestion that such a fault might be present in the Triassic floor at the line of stocks representing vents from a deep source of the magma was proposed in the Adams County, Pa., report (Stose, 1932, fig. 5) and again in the York County, Pa., report (Stose and Jonas, 1939, pp. 118-119).

According to this hypothesis the thick deposits of Triassic sediments, aggregating over 20,000 feet, would be limited to the deeper part of the basin east of the postulated fault. West of this fault the floor, composed of Paleozoic limestone, stood higher and received only the youngest Triassic sediments which overflowed the deeper basin (fig. 3). The overflow of the youngest Triasisc sediments onto the limestone floor northwest of the main basin was briefly referred to in the report on Carroll and Frederick Counties, Maryland (Stose and Stose, 1946, pp. 87-88). Movement on this fault must have ceased before the end of deposition and intrusion of diabase, as no evidence of faulting is observable in the sedimentary rocks now at the surface.

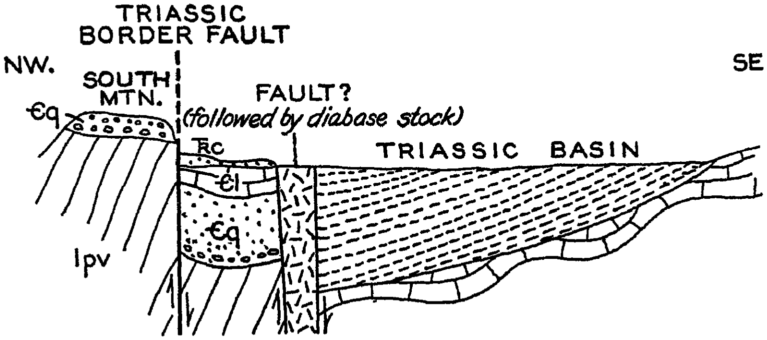

Fig 3. Section across the Triassic basin, showing the western border fault, a postulated fault in the floor of the basin followed by a diabase stock, and the elevated floor west of this fault that received the overflow from the deeper Triassic basin to the east.

The explanation here offered accounts for the exposure of Paleozoic limestone as part of the floor beneath the youngest Triassic sedimentary rocks in the western part of the Carlisle quadrangle, where the greatest thickness of sedimentary rocks should normally occur. According to this interpretation, a fault, now covered or concealed, marked the west edge of the deeper part of the Triassic basin; this fault penetrated the floor of the basin to a depth of about 20,000 feet and tapped the source of deep-seated diabasic magmas; the diabase rose into the Triassic sedimentary rocks as cross-cutting bodies and stocks, and suplied the magnetite which has been concentrated in the metamorphosed sediments; a fault along the west edge of the Triassic was active at about the same time and Paleozoic limestone, which formerly capped the volcanic rocks west of this fault, was dropped to the level of York Springs and became the floor on which the youngest Triassic sediments were deposited as overflow from the main basin; later movement on the western fault raised the Cambrian quartzite and limestone west of the fault, and waste from their erosion became the conglomerates in the closing stages of deposition; erosion followed later uplift, and removed the Cambrian limestone and quartzite from the pre-Cambrian volcanic rocks now exposed west of the fault; still later movement on this fault has been postulated (Stose, 1927) to account for the displacement of the Schooley peneplain.

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U. S. GEOLOGICAL SURVEY
WASHINGTON, D C.

FABRIC CHANGES IN YULE MARBLE AFTER DEFORMATION IN COMPRESSION

ELEANORA BLISS KNOPF

PART II

EXPERIMENTAL DEFORMATION OF YULE MARBLE

After the fabric of the undeformed marble had been studied, test cylinders were cut from the marble so that in one set, known as the P cylinders, the cylinder axis is the north-south direction, i.e., the normal to the P face of the thin section block. (See fig. 1,b). In the second set, known as R cylinders, the cylinder axis is the vertical axis in nature, i.e., the normal to the R face of the thin section block. In the third set of cylinders, known as the Q cylinders, the cylinder axis is the east-west direction, i.e., the normal to the Q face of the thin section block. Each cylinder was marked with an arrow indicating a known direction normal to the cylinder axis, with the result that the cylinder is oriented with respect to the fabric coordinates determined by preliminary study.[25] The height of each cylinder is exactly one inch and the diameter of the cross section is one half inch. The accurate orientation of these cylinders and the careful turning and finishing required for test samples in high pressure deformation is an expensive operation. The cost of preparing the cylinders for deformation was covered by a grant-in-aid from the National Research Council, which is hereby acknowledged with deep thanks.

The technique of deformation is that described by Griggs in 1936[26] and the reader is referred to this paper for a description of the experimental work, as it has been impossible for Mr. Griggs to add a section on the technique on account of his full time diversion to war work during World War II, and his preoccupation with other responsibilities in the post-war period. In the course of preparation to resume experimental deformation in the Institute of Geophysics of the University of California at Los Angeles, Professor Griggs has very kindly furnished a considerable amount of data from his unpublished notes on work done in 1936 at the Physics Laboratory of Harvard University.

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[25] Unfortunately in cutting the test cylinders it was not appreciated that *two* direction senses are required in order to establish the unique orientation of any thin section Thus although the thin sections from the original thin section block of undeformed marble were cut with unique orientation, the thin sections from the deformed cylinders were cut with an ambiguity in the sense in two out of three recorded directions. This ambiguity was eliminated as far as possible by mutual comparison of all three sections for each cylinder and by recognition of residual lamellae maxima in the deformed fabric.

[26] Griggs, David T. Deformation of rocks under high confining pressures. Jour. Geology. 44, 541-577. 1936.

PROPOSED PLAN FOR EXPERIMENTAL DEFORMATION OF YULE MARBLE

The experimental deformation of the Yule marble cylinders was planned so as to apply compressional force to the dry marble upon each of the three faces, P, Q, R, of the thin section block under definite pressure-temperature conditions and also to the wet marble in the presence of carbonated water at temperatures that would favor the operation of recrystallization in the new fabric. It was also planned to apply a tensile stress to P, Q, R cylinders under similar conditions. Finally it was hoped: (1) that after cylinders deformed under conditions described above had been studied by petrofabric analysis it would be possible to deform a new set of cylinders whose axes would be inclined to the faces P, Q, R: (2) that deformation could be stopped at various stages of deformation, 5%, 10%, etc., and that the resultant change in fabric could then be determined during the progress of deformation.

The experimental work in compression and tension on P, Q, R cylinders was carried on in the Harvard Physics Laboratory by Griggs in cooperation with his assistant, James R. Balsley, Jr.[27] Four cylinders were deformed in compression: two P cylinders, one R, and one Q cylinder. Deformation was produced under three different pressure-temperature conditions:—1. Dry deformation at 20° C.—two cylinders were used: one R cylinder, one Q cylinder; 2. Dry deformation at 150° C.—one P cylinder; 3. Deformation at 150° C in the presence of carbonated water—one P cylinder. Three cylinders, one P, one Q, and one R cylinder, were deformed in tension by Balsley. In the spring of 1941, he reported preliminary results of the deformation of the Yule marble in tension, at the meetings of the Tectonophysics Section of the American Geophysical Union[28] As the experimental deformation was necessarily suspended during the duration of the war it has been impossible to carry out further stages of the work. It is hoped that the experimental work can be resumed in the near future and in the meantime Professor F. J. Turner of the University of California has made an exhaustive fabric study of more material from the Yule quarry, preparatory to further experimental work.

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[27] This experimental work was supported jointly by the Committee on Geophysical Research and the Society of Fellows, Harvard University The apparatus for the tension tests was constructed under Grant No 201-37 made from the Penrose Research Fund of the Geological Society of America. For all this assistance sincere thanks are due.

PROPOSED PLAN FOR PETROFABRIC ANALYSIS

After removal from the testing machine it was planned to cut thin sections from the surface normal to the cylinder axis and also from two mutually perpendicular surfaces parallel to the cylinder axis, in such a way that the three thin sections would correspond to the P, Q, and R faces of the original thin section block before deformation.

The following table shows the pressure-temperature conditions under which the P, Q, R cylinders were deformed in compression and the thin sections of the deformed cylinders that were studied petrographically and by petrofabric analyses.

DEFORMATION IN COMPRESSION

Experiment 70 — (R cylinder deformed dry at room temperature).

In this experiment the cylinder was cut so that the axis was normal to the R face of the thin-section block of undeformed marble and thus corresponds to the vertical axis in nature. Therefore the cylinder axis is approximately parallel to the *s* plane of the undeformed fabric (see fig. 1, a) and makes an angle of about 26° with the preferred orientation of the calcite optic axes.

The marble was jacketed in copper foil and deformed at room temperature (20°C) under a confining pressure of 10,000 atmospheres. Total deformation was about 36%, of which 16% was recovered on release of stress and removal

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[28] Balsley, James R. Deformation of Marble under Tension at High Pressure. National Research Council Transactions, American Geophysical Union, Pt 2, 519-525, 1941.

Experimental Work on Yule Marble in Compression

By D. T. Griggs, in the Physics Laboratory,
Harvard University, 1936.

Cylinders cut with axis parallel to vertical axis in nature are R.
Cylinders cut with axis parallel to N-S axis in nature are P.
Cylinders cut with axis parallel to E-W axis in nature are Q.
Strength at atmospheric pressure = 4,220 lbs./sq in.[29] (measured on R cylinder).

Exp.	Cylinder Axis	Confining Pressure in Atmospheres	Strain Meas'd Under Pressure	Recovery	Permanent Deformation	Change in Sp. Gr.	Symmetry of Deformation
70R	Vertical	10,000[30] t = 20°C	36%	16%	20%	—11%	Orthorhombic
73Q	E-W	10,000 t = 20°C	45%	20%	25%	—8%	Axial
74P	N-S	10,000 t = 150°C in the presence of carbonated water	39.3%[31]	(—1 3)%[31]	40.6%		Orthorhombic
75P	N-S	10,000 t = 150° dry	~30%[31]	(~2 4)%[31]	27.6%	—~9%	Orthorhombic

	Remarks
70R	Jacketed in Cu foil; load release instantaneous; elliptical cross section; ratio 1:1 20.
73Q	Jacketed in Cu foil; jacket broke; spec. broke on vertical N-S plane; circular cross section. Ellipticity 1.01.
74P	Jacketed in Cu foil; jacket broke, elliptical cross section; minor axis of ellipse parallel to dip; deforms by jerks.
75P	Jacketed in Cu foil; elliptical cross section; ratio 1 1.23; minor axis of ellipse parallel to the dip

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[29] For strength to rupture on variously oriented specimens see Lepper, Jr., H. A Compression tests on oriented specimens of Yule marble, in this number of Am Journal of Science.

[30] ± 10%.

[31] Apparatus not accurately calibrated at 150°C

from the apparatus, so that the permanent deformation amounted to 20%.[32] As far as could be judged from the appearance of the test piece, deformation was homogeneous, with no measurable bulge. The cross section of the test piece became an ellipse having a ratio of 5:6 between the minor and major axes. The major axis lies in the east-west direction of the undeformed marble, thus indicating greater freedom of movement of material in the east-west than in the north-south direction, in other words, an orthorhombic rather than an axial strain.

Strength of the material at the maximum, when the experiment terminated, was more than 160,500 psi (see fig. 48) and the specific gravity was reduced 11%, measured after removal from the apparatus.

PETROFABRIC ANALYSIS

Thin sections were cut from the deformed test specimen as follows· one parallel to the end of the test pieces, which is parallel to the R face of the original thin section block; one parallel to the P face; and one parallel to the Q face of the original block. (See fig. 1,b). These sections were labelled 70R, 70P, and 70Q respectively. Photomicrographs of these thin sections are shown in plates 5, 6, 7.

A pronounced dimensional anisotropy in the grains can be recognized in the thin section, and is completely reoriented from the dimensional arrangement in the undeformed fabric. The P section shows a tendency towards grain elongation in the east-west direction; average grain size in the east-west direction is .75 mm. The Q section shows elongation in the north-south direction; average for the long diameter is .9 mm. In the R section the grains are more nearly equidimensional; average grain diameter .6 mm. This means that relatively flat grains are arranged with their flat sides in the horizontal plane, as contrasted to the undeformed fabric, in which the flat side of the grains occupies the north-south vertical plane. In the flat grains there are one or two sets of closely spaced lamellae. These lamellae are more numerous and more closely spaced than in the undeformed fabric and in many places are bent or twisted. Many grains are cloudy. They are more conspicuous in the sections cut parallel to the axis of compressive force than in the section normal to the applied force.

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[32] In this experiment owing to a failure in the upper piston, the confining pressure was released instantaneously without, however, causing any recognizable damage to the specimen.

Complete collective diagrams were prepared from the calcite lamellae and axes in the P, R, and Q sections. The same technique was used as in measuring the undeformed fabric. Grains numbered in the photomicrograph were measured and recorded, grain by grain. Measurements were plotted and contoured in elemental diagrams to test the homogeneity of the fabric in the thin section. The elemental diagrams show a marked similarity, which indicates an essential homogeneity of fabric, although there is a slight inhomogeneity between P and Q.

LAMELLAE DIAGRAMS

P Section: An elemental diagram from the P section (fig. 24) shows a marked concentration of lamellae poles lying close to the horizontal plane. A collective diagram made from 281 poles of calcite lamellae (fig. 25) shows the position of the main concentration of lamellae and indicates that this concentration is distributed into a maximum A and submaximum B, occupying two intersecting planes that make angles of 20° each with the horizontal plane. There is a stronger concentration of lamellae in the plane that dips 20° west (maximum A) than in the plane dipping 20° east (submaximum B), indicating an intensification of movement along one of the intersecting planes. A submaximum C about 25° from the east-west pole of the diagram occupying a plane that dips steeply east, probably represents a relic of the lamellae maxima of the undeformed fabric. The collective diagram of

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In the following diagrams the one known direction sense in each thin section has been underlined

Figure 24 50 poles of calcite lamellae from P section of R cylinder after deformation. Contours 8-6-4-2-0%.

Figure 25 281 poles of calcite lamellae from P section of R cylinder after deformation. Contours (7-5)-3-2-1-0%.

Figure 26. 320 poles of calcite lamellae from P section of R cylinder after deformation including 56 poles rotated from R to fill P Contours (4-3)-2 5-2-1-0%.

Figure 27 61 poles of calcite lamellae from R section of R cylinder after deformation. Contours (9-8)-6-4-3-2-0%.

Figure 28 146 poles of calcite lamellae from R section of R cylinder after deformation Contours (7-6)-4-3-2-1-0%

Figure 29. 305 poles of calcite lamellae from R section of R cylinder after deformation· 150 poles measured in the R section, 155 poles rotated from P section Contours (6-5)-4-3-2-1-0%.

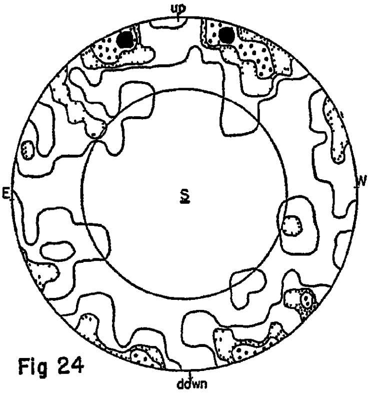

Fig 24

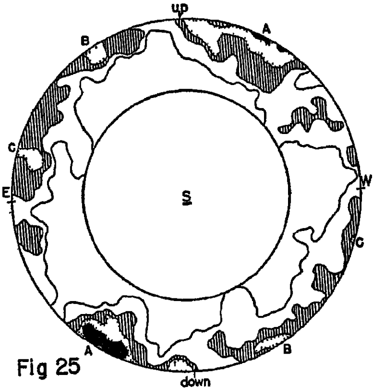

Fig 25

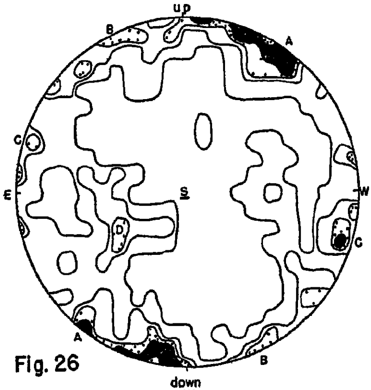

Fig. 26

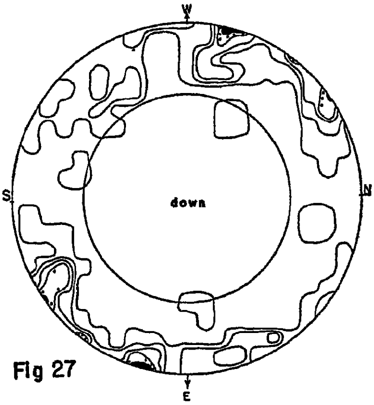

Fig 27

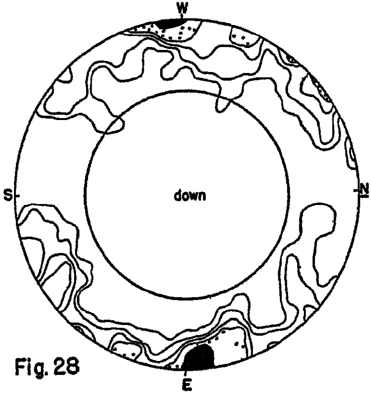

Fig. 28

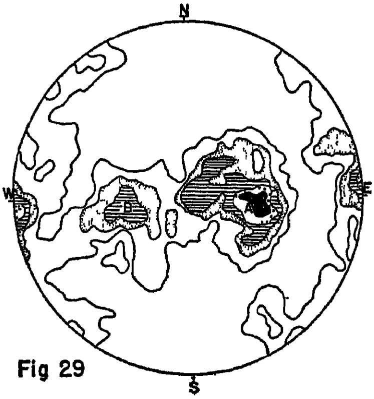

Fig 29

P, filled from the R section, (fig. 26) indicates a minimum around the north-south direction and shows some concentration of lamellae (D) in the position occupied by the K maximum in the undeformed fabric. This submaximum at D probably is another relic of the lamellae maxima of the undeformed fabric.

R Section: An elemental diagram of 61 lamellae was contoured from the R section (fig. 27). Two other elemental diagrams of 50 and 40 lamellae respectively, established that the fabric is essentially homogeneous. Maxima are scattered and not very definite in position or outline, which is expectable because the predominant maxima from the P section lie close to the axis of compression and would therefore be unmeasurable in the R section. The maxima in R occupy the periphery of the diagram in a fairly wide arc on both sides of the emergence of the east-west axis. The most definite feature of the three elemental diagrams is the distinct minimum around the emergence of the north-south axis. The fabric as determined from figure 28, which is an unfilled collective diagram of 146 lamellae poles in R is not essentially unlike that in the undeformed marble (fig. 16), confirming the conclusion that this diagram represents a relic fabric. The new fabric associated with the new position of the megascopic slip plane that is indicated by the dimensional arrangement of the grains could only appear in diagrams measured from P or Q. In the *complete* collective diagram of R filled from P (fig. 29) it shows plainly.

Q Section In the collective diagram measured from the Q section the lamellae are concentrated in a peripheral girdle

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Figure 30 380 poles of calcite lamellae from Q section of R cylinder after deformation: 257 poles measured in Q section, 123 poles rotated from P section Contours > 5-4-3-2-1-0%

Figure 31 80 poles of calcite lamellae from upper end of R cylinder after deformation, measured in Q section. Contours (12-10)-8-5-3-0%.

Figure 32. 68 poles of calcite lamellae from middle of R cylinder after deformation; measured in Q section Contours (10-9)-8-6-4-3-0%.

Figure 33 281 poles of calcite lamellae from P section of R cylinder after deformation; rotated to position of Q for comparison with Figure 30. Contours (7-5)-3-2-0%

Figure 34. 53 optic axes of calcite from R section of R cylinder after deformation. Contours 8-6-4-0%.

Figure 35 70 optic axes of calcite from Q section of R cylinder after deformation. Contours (10-8)-6-3-0-%.

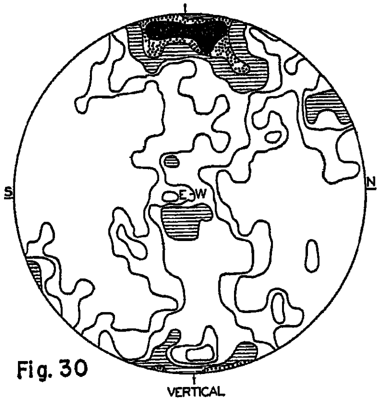

Fig. 30

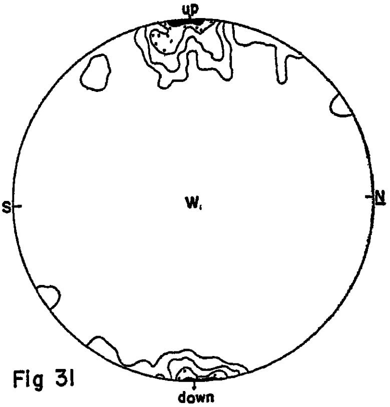

Fig 31

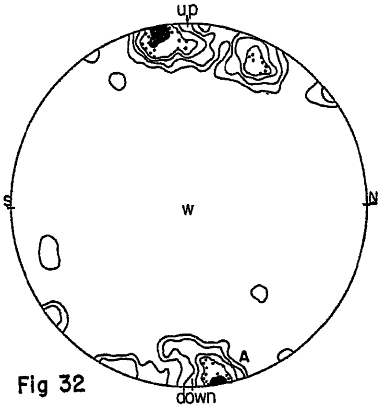

Fig 32

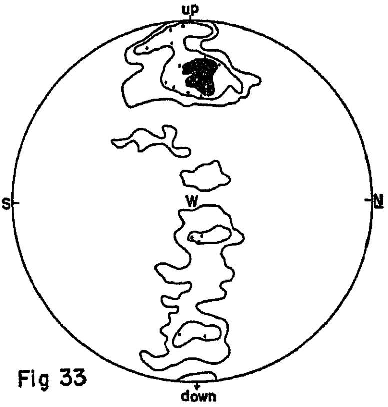

Fig 33

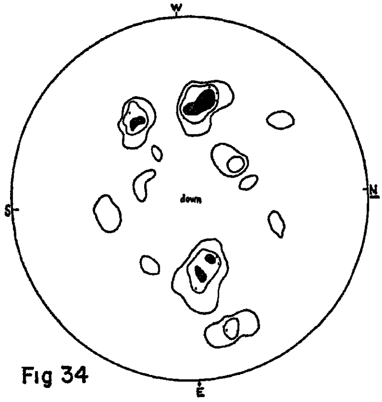

Fig 34

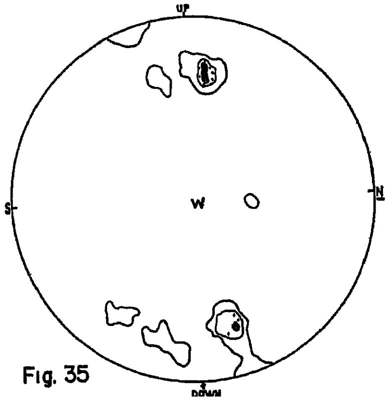

Fig. 35

with the maximum (fig. 30) close to the emergence of the vertical axis. There is a small angular divergence amounting to about 10° between the lamellae maximum and the vertical axis of the cylinder and the maximum is elongated in the E-W direction in the section measured from Q. In order to find out whether there is an essential difference between the course of the deformation at the end or in the middle of the cylinder a diagram was made from 80 lamellae at the upper end of the cylinder (fig. 31) and another from 63 lamellae from the middle of the cylinder (fig. 32). These two diagrams are essentially the same, showing that the deformation was penetrative and almost uniform in operation throughout the cylinder. In the complete collective diagram of 380 lamellae, 257 measured in Q and 123 rotated from P (fig. 30), the distance between the pole of the vertical cylinder axis and the lamellae maximum is considerably less in the Q than in the P section (fig. 25), so much so that even allowing for a possible divergence of a few degrees from 90° between the two sections due to inaccuracy of cut it may indicate a slight inhomogeneity in the fabric corresponding to the small rotation around the N-S axis of the original fabric. In order to test the degree of homogeneity in the fabric, the diagram of 281 lamellae measured in P (fig. 25) was rotated to the position of Q (fig. 33) to compare with the diagram of 380 lamellae in figure 30. The general correspondence of the pattern in the two diagrams outside of the blind area in Q is reasonably good.

AXES DIAGRAMS

Measurement of axes in these thin sections proved to be very unsatisfactory because: 1, the thin sections cut from the deformed cylinders are thicker than the standard thin section; 2, many calcite grains are twinned repeatedly, so that the twin lamellae are very closely spaced and do not allow the determination of axes; and 3, most grains are slightly biaxial, the axial angle being small and variable in amount.

It was possible to measure from the R section 53 axes (fig. 34), from the Q section 70 axes (fig. 35), and from the P section (fig. 36) 57 axes with an approximate degree of accuracy. The diagrams from the R and Q sections coincide reasonably well. In both diagrams two maxima lie on the vertical great circle passing through the east-west direction. The maxima are about 35° distant from the axis of compressive force, which is, in this cylinder, the vertical direction.

In the R section the axes and lamellae maxima lie upon the vertical great circle passing through the east-west direction, thereby establishing the position of the deformation plane and the direction of movement as east-west, a conclusion that is confirmed by the maximum extension of the deformed cylinder in the east-west direction.

A *slight tendency to elliptical arrangement of the axes* in the R section coincides with the orientation of the elliptical cross section of the deformed cylinder and may suggest the influence of an elliptic conoid of shear, the major axis being in the east-west direction. However, this elliptical arrangement is not noticeable in the Q section, although the two axes maxima in the Q section lie upon the trace of this ellipse in Q.

In the P section one axis maximum lies so near to the position of the axes in the undeformed fabric that it probably represents a residual from the undeformed fabric. A maximum and a submaximum near the vertical poles probably represent changes of positions of axes during the course of deformation.

In order to estimate the relative frequency of various crystal planes among the measured lamellae, 118 lamellae were tabulated from the Q section of the deformed R cylinder. The tabulation showed: (1) the angles between the lamellae poles and the optic axes, (2) the angles between the lamellae poles in grains where more than one set of lamellae were measured. The following table shows the values for the crystallographic angles that occur most often.

Values of crystallographic angles in calcite

cr	$(0001) \wedge (10\bar{1}1)$	=	44°36½′
ce	$(0001) \wedge (10\bar{1}2)$	=	26°15′
rr	$(10\bar{1}1) \wedge (\bar{1}101)$	=	74°55′
ee	$(01\bar{1}2) \wedge (\bar{1}012)$	=	45°3′
cf	$(0001) \wedge (02\bar{2}1)$	=	63°7′
cm	$(0001) \wedge (10\bar{1}0)$	=	90°

Order of frequency of crystal planes among measured lamellae (in per cent of total grains measured)

Grains measured	Crystal plane	Frequency of occurrence
118	*e* $\{01\bar{1}2\}$	25%
	r $\{01\bar{1}1\}$	15%
	m $\{10\bar{1}0\}$	17%
	f $\{02\bar{2}1\}$	10%

However, it is impossible to measure enough optic axes with sufficient accuracy to justify any conclusion about the behavior of individual grains and the interpretations based upon the few measurements that were available can only be considered tentative unless confirmed by later work.

Experiment 75— (P cylinder deformed dry at 150°C.)

In this experiment the axis of the cylinder was cut parallel to the north-south axis in nature, i.e., normal to the P face of of the thin-section block in the undeformed marble. Therefore in this experiment as in 70 R the cylinder axis is nearly parallel to the plane of the grain (*s* plane) in the original fabric. (See fig. 1,a). The marble was jacketed as in Experiment 70 and deformed under a confining pressure of 10,000 atmospheres, the principal difference between the deformation of this cylinder and that of the R cylinder being that in this experiment the temperature was raised to 150° C. The permanent deformation in this experiment was 27.6%. As in Experiment 70 deformation was homogeneous, as far as could be determined from the external appearance of the test cylinder. The cross section of the test piece was again an ellipse with the major axis again oriented in the east-west direction.

PETROFABRIC ANALYSIS

Thin sections were cut from the P test cylinder just as in the R cylinder; one, parallel to the end of the cylinder is also parallel to the P section in the undeformed fabric, two other mutually perpendicular sections are parallel to the cylinder axis and to the R and Q section in the undeformed fabric (see fig. 1, b). Photomicrographs of these sections are shown in plates 8, 9, 10.

The three sections show the same dimensional anisotropy in the grains of the deformed marble, as seen in Experiment 70. In this cylinder, however, elongation is in the vertical direction in the Q section and in the E-W direction in the R section. In the P section the grains are more nearly equidimensional. In other words, the elongation represents a flattening in the directions normal to the axis of compression just as in Experiment 70. In the thin section the large grains are surrounded in some places by a border of smaller grains. The larger grains have clear-cut distinct outlines, forming a mosaic. The traces of the lamellae in the deformed grains are much closer spaced and more abundant than in the fabric of the orginal Yule marble. In the Q and R sections almost every grain shows one or two sets of closely spaced lamellae. In the P section (normal to the compression axis) a number of grains are relatively free from lamellae.

The thin sections were studied in the same way as the thin sections of the R cylinder. Measurements were recorded, then plotted and contoured, first in elemental, then in collective *diagrams, and finally the lamellae diagrams were filled by* rotation.

LAMELLAE DIAGRAMS

Q Section 280 lamellae were measured in 207 grains. These were contoured as 73 lamellae from 52 grains, 170 lamellae from 100 grains, and a final collective diagram of 280 lamellae from 207 grains. These diagrams were so consistently uniform, indicating such an essential homogeneity of the fabric, that only the final diagram (fig. 37) is published. The strong maximum A centers around the ends of the north-south pole in nature, which is the axis of compressive force in the experiment, and a moderately strong submaximum B occurs at about 24° below the N-S pole.

A diagram from 64 lamellae poles measured from the end of the cylinder is essentially homogeneous with a diagram made from 76 lamellae poles measured in the middle of the cylinder, thus showing, just as in Experiment 70, that the deformation was a homogeneous strain throughout the whole cylinder.

P Section: A diagram was made from 115 lamellae poles measured in the P section (fig. 38). This shows almost a complete peripheral girdle with a distinct minimum around the vertical direction in nature. The maxima that appear on either side of the east-west axis of the diagram are probably residual from the undeformed fabric.

The prominent maximum that appears in the Q section, which is the new arrangement induced by the deformation, could not of course appear in the unfilled diagram of P as it would occupy the blind spot.

R Section: A diagram of 63 lamellae, measured in the R section (fig. 39) shows an area of maximum concentration about 35° west of north and another maximum about 15° west of north. These are clearly newly developed concentrations resulting from the compression in a north-south direction. There is a submaximum B in the northwest-southeast quadrant of the diagram at almost 80° from the dominant maxima at A. This submaximum appears with varying intensity in the elemental diagrams of 63 and 70 lamellae respectively, and probably represents a residual concentration from the original fabric before deformation.

The collective diagram of 135 lamellae (fig. 40), which corresponds perfectly to the elemental diagrams, shows the new maxima extending through an arc of about 50° west of north close to the axis of compression. This resembles so closely the arc of about 50° close to the axis of compression in Experiment 70 that it probably indicates the same flattening of grains in a plane normal to the axis of compression that is shown in Experiment 70. This same flattening of grains in the deformed P cylinder can be seen plainly in the photomicrographs of the R section (pl. 9) and of the Q section (pl. 10). The extension in the east-west direction shown in plate 9 is greater than in the vertical direction in plate 10. This greater east-west extension conforms to the east-west major axis of the elliptical cross section of the P cylinder.

The asymmetric position of the new maximum on the west side of the north-south axis in the R section shows a certain lack of homogeneity between the asymmetry in the R and in the Q sections that might be explained by a slight error of orientation in cutting the R section.

AXES DIAGRAMS

The same difficulty as in Experiment 70 was found in making accurate determination of optic axes orientation in Experi-

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Figure 36. 57 optic axes of calcite from P section of R cylinder after deformation Contours (12-10)-7-3-0%.

Figure 37. 280 poles of calcite lamellae from Q section of P cylinder after deformation. Contours (12-10)-8-6-4-2%.

Figure 38 115 poles of calcite lamellae from P section of P cylinder after deformation Contours (7-5)-4-3-2-0%

Figure 39. 63 poles of calcite lamellae from R section of P cylinder after deformation. Contours (10-7)-6-5-3-0%.

Figure 40 135 poles of calcite lamellae in R section of P cylinder after deformation Contours (7-5)-4-3-2-0%

Figure 41. 62 optic axes of calcite from R section of P cylinder after deformation. Contours > 10-8-4-0%.

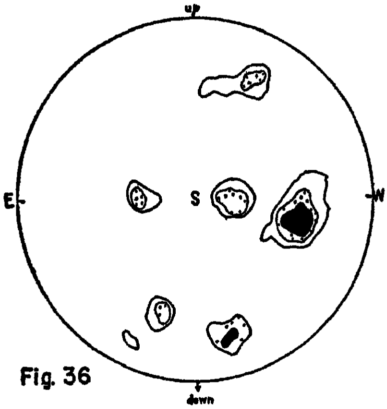

Fig. 36

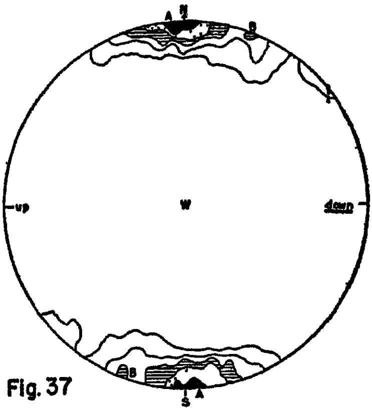

Fig. 37

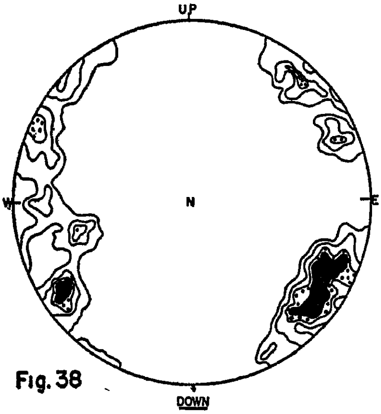

Fig. 38

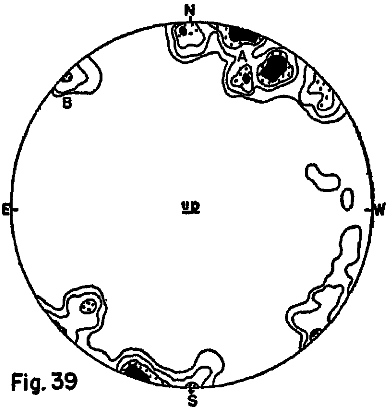

Fig. 39

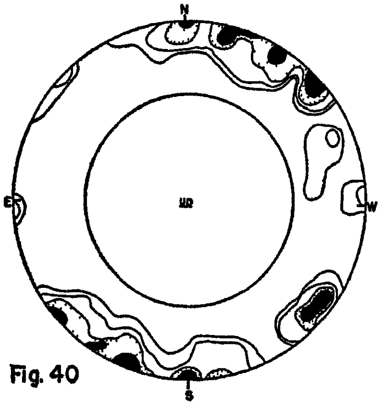

Fig. 40

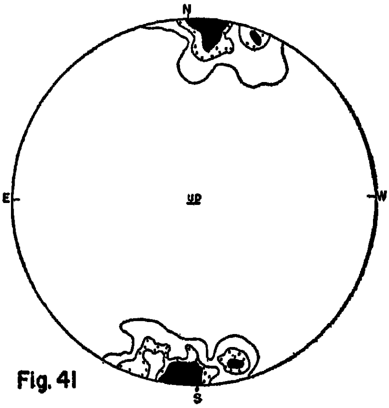

Fig. 41

ment 75. In the R section satisfactory measurements of 55 grains were obtained after checking each measurement three times. The significance of the pattern was established by the fact that its essential features were shown in a preliminary count of as few as 30 optic axes. The most that could be definitely established is that the optic axes in the R section of the P cylinder (fig. 41) are near the N-S direction in nature, thus near the lamellae pole maximum of the R section. This conclusion is justified by the general homogeneity of fabric indicated by the 62 optic axes measured in the R section (fig. 41) when compared with a collective diagram of 188 optic axes measured from the Q section (fig. 42). The same homogeneity of fabric was demonstrated by rotating the 62 axes measured in R (fig. 41) to the position of P for comparison with figure 43, which is a diagram of 163 axes in the P section of the P cylinder, measured independently at Harvard University in 1938.

The essential character of the diagram of 188 axes from the Q section (fig. 42) is clearly foreshadowed in elemental diagrams made from 52 axes, and 102 axes. In all these diagrams the two maxima are clustered at and near to the north-south direction, which is the axis of compression in the experiment.

This coincidence in pattern shows that the diagram in figure 42 is really a significant fabric pattern in spite of the difficulty in measuring optic axes in thick sections of strongly deformed calcite.

The biaxility of the calcite in P was examined in sodium light in thirty five grains. Of these about one half are clearly biaxial. Biaxial grains give no sharp extinction but show a

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Figure 42. 188 optic axes of calcite from Q section of P cylinder after deformation. Contours (8-5)-4-3-2-0%.

Figure 43. 163 optic axes of calcite from P section of P cylinder Contours > 9-7-5-3-1-0%. (Diagram made by Snively and Bell.)

Figure 45. Cyclographic projection of a calcite crystal showing the areas where twin-gliding on one glide plane is impossible. Adapted from a diagram by James Forbes Bell.

Figure 46 Diagram showing areas in a calcite crystal where twin-gliding on $\{01\bar{1}2\}$ under compression parallel to the optic axis is impossible Adapted from a diagram by James F Bell

Figure 47. Diagram showing fields of permissible twin-gliding in undeformed Yule marble.

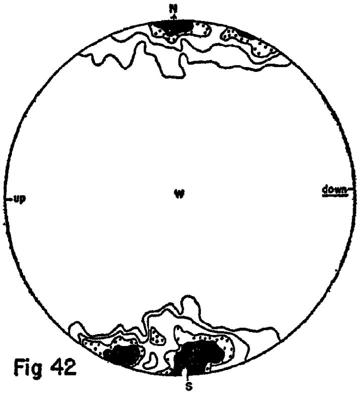

Fig 42

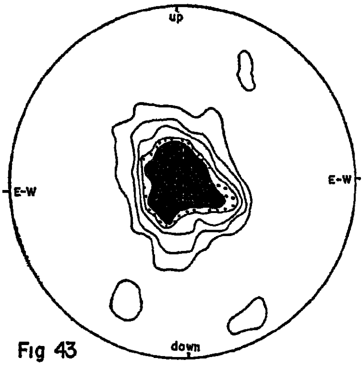

Fig 43

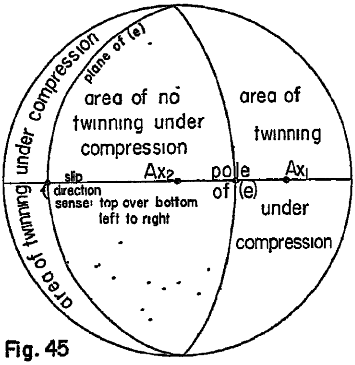
plane of (e)
area of twinning under compression
area of no twinning under compression
area of twinning
slip direction
sense: top over bottom left to right
Ax2
pole of (e)
Ax1
under compression

Fig. 45

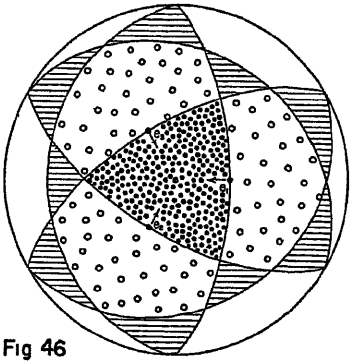

Fig 46

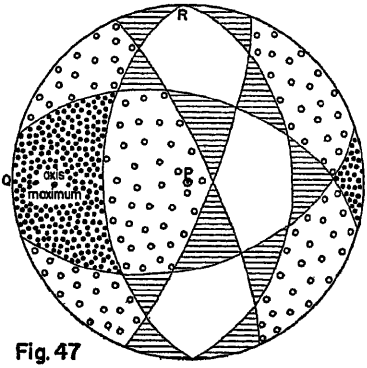

Fig. 47

blue mottling in the extinction position, a fact that has been noted by Gillson.[33]

Calcite grains in the P cylinder are much less murky than in the R cylinder and it is easier to get a good interference figure in sections normal to the cylinder axis than it is in sections normal to the axis of the R cylinder, therefore the biaxiality is more easily recognized in the P cylinder. In convergent light, fourteen out of eighteen biaxial grains in R showed a very small axial angle. No satisfactory estimate could be made for the percentage of biaxial grains in the R cylinder.

The relative frequency of occurrence of various crystal planes among the measured lamellae was examined in the same way as in Experiment 70 and the following table gives the results.

Order of frequency of crystal planes among measured lamellae (in percent of total grains measured)

Grains measured	Crystal plane	Frequency of occurrence
286	*e* $\{01\bar{1}2\}$	40%
	r $\{01\bar{1}1\}$	13%
	m $\{10\bar{1}0\}$	$<10\%$
	f $\{02\bar{2}1\}$	12%
	c $\{0001\}$	$<6\%$

The most striking difference between the deformed R and P cylinders is that the *e* plane is more strongly developed in the P cylinder. Also the basal plane appears to have functioned to a limited extent in the gliding process in P, whereas in the R cylinder there is no evidence of gliding on the base. It was pointed out by Eskola[34] that the structure of calcite can be considered as a distortion of the halite cubic structure. By this distortion the cubic planes would become the rhombohedral planes *r*. The plane that truncates the three rhombohedral edges in calcite is the basal plane. The plane that truncates three cubic edges in halite is the octohedron. It has been shown by Tammann and Salge[35] that above 100°C halite glides on (111) in a direction parallel to the edge [001:110]. It is interesting to note that in calcite the corresponding direction would be the direction in (0001) that is parallel to the edge $[01\bar{1}1:0001]$, which is the direction of alinement of nearest closest like ions, and therefore theoretically an expectable translation direction.[36]

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[33] Gillson, J. L. Biaxial calcite. Am. Mineralogist, 12, 357, 1927.

[34] Eskola, P. Kristalle und Gesteine. p 185. Springer Verlag. Wien 1946.

[35] Tammann, G. and Salge, W. Neues Jb. B. B. 57 A. 117, 1928.

Experiment 74 P — (P cylinder deformed at 150° C in carbonated water).

In this experiment the cylinder axis was parallel to the north-south direction in nature, normal to the P face of the thin-section block in the undeformed marble. The cylinder axis, as in 75 P, is nearly parallel to *s* in the original fabric.

The marble was jacketed in copper foil and deformed under a confining pressure of 10,000 atmospheres at 150°C in the presence of commercial carbonated water. Permanent deformation was 40.6%, the highest attained in any experiment. The deformation in this experiment took place by jerks. As in 70 R and 75 P the strain was orthorhombic and the deformed test piece again showed an elliptical cross section with the major axis of the ellipse in the east-west direction.

Extremely interesting is the texture of the deformed marble shown in the photomicrograph in plate 11. It is entirely different from the texture of cylinders that were deformed dry. The grains are clear, not turbid. Lamellae are sharp and straight with no sign of deformation and the texture is clearly that of a recrystallized fabric.

Unfortunately only one thin section was cut from this cylinder and study of this section has raised some doubt as to the correct orientation of the section. Therefore it has been impossible to draw any reliable conclusions about the deformed fabric. It is hoped that this extremely interesting experiment will be repeated soon and an exhaustive petrofabric study of the deformed material be made.

Experiment 73 Q — (Q cylinder deformed dry at room temperature).

In this experiment the axis of the cylinder was parallel to the east-west direction in nature, normal to the Q face of the thin-section block in the undeformed marble. Therefore the cylinder axis is normal to *s* in the undeformed fabric (see fig. 1,a) and makes an angle of about 26° with the optic axis maximum.

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[36] Buerger, M. J. Translation-gliding in crystals. Am. Mineralogist, 15, 46, 1930.

The cylinder was jacketed in copper foil and the experiment was carried on at room temperature under 10,000 atmospheres confining pressure. During the run the jacket broke, permitting access of kerosene, and the specimen broke on the vertical plane striking north. Total deformation was 45%; recovery was 20%, leaving 25% permanent deformation. Reduction in specific gravity was 8%. The east-west extension attained in the other experiments was prohibited in this cylinder by the orientation and the deformation was an axial strain, producing a circular cross-section in the deformed cylinder.

Unfortunately no petrofabric analysis could be made, as the cylinder crumbled to powder before it could be sectioned.

RELATION OF DEFORMED TO UNDEFORMED FABRIC

The deformed fabric in the dry deformation of marble is characterized by the following features:

1. The change of grain shape from a position in the undeformed marble where the ellipsoidal grains are arranged with their shortest axis parallel to the east-west direction in nature to a position in the deformed fabric where the shortest axis of the ellipsoidal grains lies parallel to the axis of compression, regardless of whether the compressive force was applied parallel to the north-south or to the vertical direction in the undeformed fabric.

2. The lamellae in the deformed fabric are much more abundant and closer set than in the original fabric — they are frequently bent.

3. Biaxial character of calcite is more pronounced than in the original fabric.

4 The change in position of the ellipsoidal grains is accompanied by a general migration of lamellae maxima from positions close to east-west in the undeformed fabric to positions near the vertical in the deformed fabric of the R cylinder or near the N-S direction in the deformed fabric of the P cylinder, (fig. 1,b) i.e., to positions nearly normal to the axis of compression. This change is accompanied by a similar migration of optic axes from an almost horizontal position, a little north of west, to a position near the axis of compression. The shift in preferred orientation suggests

that the flattening of calcite grains in a plane normal to the compression has been accomplished by a mechanism of intragranular slip.

5. The symmetry of the deforming movement is not axial as might be expected from the operation of a cone of shear in isotropic material but is orthorhombic giving rise to the elliptical cross section of the deformed cylinder. The position of the major axis of this ellipse, i.e., the east-west direction, is predetermined by the relation of the optic axis maximum in the undeformed fabric to those lamellae maxima that make 26° angles with the optic axis maximum.

MECHANISM OF SLIP

It has been a matter of observation for many years that calcite can be made to change its optical orientation by twinning. Baumhauer[37] demonstrated this fact by pressing with a knife blade upon the edge [r:r]. The resultant crystal appears as in figure 44,a, which is slightly modified from a figure published by Denaeyer.[38]

The optic axis in the twinned portion of the crystal has moved through an angle of 52° from the original position. The movement was conducted along the glide plane e (01$\bar{1}$2) in a direction that is fixed by the intersection between the glide plane and that plane normal to the glide plane which contains the optic axis of the untwinned portion of the crystal and the normal to the glide plane (fig. 44,b). This plane normal to the glide plane is the deformation plane of the twin-gliding movement.

It is obvious that a pressure P_2 in figure 44,a, applied upon the twin-gliding plane from the direction into which the twinned part of the crystal must move cannot produce twinning, because twin-gliding is a polar operation that can only be carried out in the direction sense required by the movement of the original optic axis through the pole of the glide plane into its new position.

Translation-gliding, which is a nonpolar movement, involving no change in optical orientation but merely a change in shape similar to slip of cards in a card deck, can take place in a direction sense that is prohibited for twin-gliding. Thus although pressure applied in the direction P_2 cannot produce

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[37] Baumhauer, H Über kunstliche Kalkspath-Zwillinge nach—1/2R—Zeitsch Krist., 3, 588, 1879

[38] Denaeyer, M E Caractères optiques des lamelles hémitropes b' (01$\bar{1}$2) dans les clivages minces de la calcite. Soc Géol Belgique, 64, Bull. 1, 2. 57-63, 1940

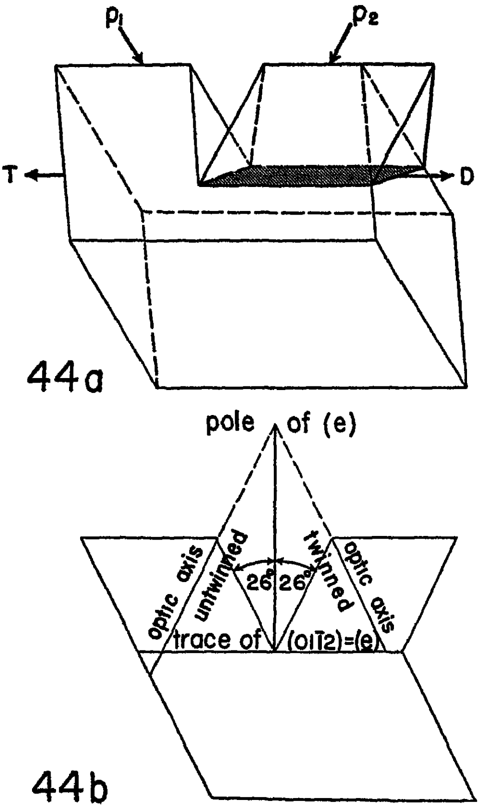

Figure 44. Diagram to show action of twin-gliding in calcite (not drawn to scale).

(a) Adapted from M. E. Denaeyer. Shaded plane is the glide plane $(01\bar{1}2)$ D is the direction sense of the twin-gliding. The line joining DT is the direction of translation-gliding.

(b) Deformation plane of calcite twin. Position of optic axes is diagrammatic.

twin-gliding in the direction sense D it could produce translation-gliding in the direction sense T.

OPERATION OF SLIP THROUGH A DEFORMING MASS

Single crystals. It has been established for metal crystals in plastic deformation by twin-gliding or by translation-gliding, that the slip movement starts on whichever potential gliding system the resolved shear stress is a maximum (Law of maximum resolved shear stress). A critical resolved shearing stress is necessary to initiate deformation. As a result of this slip a rotation of the slip system ensues, changing the stress field on the glide plane in such a way that a second glide system can function in crystals possessing more than one potential glide systems. When the resolved shear stress on the first glide system is equal to that in the second plane, double slip takes places in such a way that the movement proceeds down a symmetry plane between the two glide systems. When the change in shape brings the long axis of the deformed crystal into a plane that contains both glide directions, the reorientation ceases. In the course of deformation the operative glide systems change their positions with reference to the applied force. They tend to flatten out by tipping in towards the axis of tensile force or away from the axis of compression force. When the glide system is almost normal to the axis of compression, or almost parallel to the axis of tension (within 19° to 20°) the deformation ceases.

Polycrystalline aggregate. Conditions that govern individual grain gliding are complicated in polycrystalline material by various factors that do not enter into single crystal deformation. The influence of grain boundaries and the relative orientations of adjacent grains are two factors that constrain the flow of individual grains in an aggregate. This fact has been demonstrated by the contrast in behavior of hexagonal magnesium, with a single slip system, and face-centered cubic aluminum with its 12 equivalent glide systems. Polycrystalline magnesium is much stronger and less easily deformed than a single crystal of magnesium because of the resistance to movement offered by the diverse orientations of the one possible glide mechanism in the single crystal. In polycrystalline aluminum the resistance to deformation is much less than in magnesium because the 12 glide systems offer much more possibility for movement in individual grains.

However, even for metal working, which has been intensively studied in the last few decades, the understanding of the mechanism of flow in polycrystalline material is still far from satisfactory. It has been established by study of deformed material that polycrystalline metal develops, by working a definite and characteristic "texture",[39] i.e., preferred orientation, just as in calcite marble deformation, calcite develops a characteristic preferred orientation pattern. How this reorientation of individual grains comes to pass is still a matter for study.

Two hypotheses that attempt to explain polycrystalline deformation on the basis of the mechanism of gliding in individual grains are:

1. Wever and W. E. Schmid's hypothesis of the actual rotation of the active translation plane around an axis that is normal to the direction of movement, — in petrofabric terms, a B axis of external rotation. Thus the ultimate position of the space lattice can be attained by the activation of two glide systems. This idea, however, fails to give a satisfactory explanation for the texture of face-centered metals deformed by rolling.

2. The explanation favored by Polanyi and other German metallographers, also by Taylor in Great Britain, was that the change in form in worked polycrystalline metal could be produced by the activation of three or, more easily of five crystallographically equivalent glide systems. Such a slip mechanism would develop a preferred orientation of glide systems normal to the axis of compressive force or parallel to the axis of tensile force.

So far very little is known about the actual mechanism of slip in polycrystalline monomineral aggregates of ionic crystals, or about the relation of twin-gliding to translation-gliding in plastic deformation of minerals.

An approach to this problem was made by Griggs,[40] who deformed clear crystals of Iceland spar free from cleavage flaws and visible twinning. Deformation was carried on at confining pressures ranging between 1 and 10,000 atmospheres. The cylinder was cut so that the long axis is parallel to one edge [r:r]. In this position two of the three potential *e* glide planes could function either by twin-gliding or translation-gliding. The third plane, being parallel to the direction of applied force, could not function. The two operative planes could conduct the extension of the crystal in a direc-

[39] "*Texture*" is used here in the sense of the metallographer, i.e. a preferred orientation pattern

[40] Griggs, David T. Deformation of single calcite crystals under high confining pressures Am. Mineralogist, 23, 28-33, 1938.

tion that would be a compromise between the two twin-gliding directions in the two operative *e* planes. As pointed out by Griggs[41], fabric analysis proves the operation of twinning on these two planes. Moreover, the cross section of the deformed calcite cylinder is an elliptical section, just as in the deformed marble cylinders. By plotting on an equiareal net the position of operative glide directions in operative twin planes shown in thin sections of the calcite cylinder, it can be demonstrated that the compromise direction between these two positions coincides with the major axis of the elliptical cross section.

It is interesting to note the increase of ultimate strength that is produced in a single crystal of calcite by raising the confining pressure to 10,000 atmospheres. However the single crystal is weaker than the polycrystalline aggregate, i.e. it deforms to a greater extent than Yule marble does under identical confining and differential pressures. On the other hand, its deformability is much less than in the polycrystalline marble because at 10,000 atmospheres confining pressure the single calcite crystal ruptured after a 7% deformation whereas the polycrystalline Yule marble under the same conditions attained a permanent strain of 20%.

Griggs[42] has noted that pure translation, if present, was a minor factor in the deformation. Possibly the fact that polycrystalline marble deforms more than a single crystal of calcite before rupture is partly explicable by the combined influence of translation and twin-gliding in the polycrystalline aggregate.

When a calcite crystal is deformed by applying a compressive force parallel to the crystallographic *c* axis, the resolved shear stress would permit translation on all three glide systems but would not permit twinning, because movement in twinning would be opposed to the compressive force. It remains for future experimental work to demonstrate the relative strength of calcite crystals in various orientations.

RELATION OF TWIN-GLIDING TO TRANSLATION-GLIDING IN DEFORMATION OF YULE MARBLE

From the present study of deformed Yule marble it is clear that intragranular slip is responsible for much of the defor-

[41] Loc cit p. 31

[42] Griggs, loc cit., p 33

mation. The complete change of arrangement of the ellipsoidal or discoid grains between the undeformed and the deformed fabric has been accompanied by such a development of new and close-set lamellar structure that intragranular gliding must have operated to a marked extent, consequently the new fabric cannot be interpreted as wholly or largely caused by an intergranular rotation of the individual grains into a new position. Moreover, the demonstrable relation of the plane, determined by the pole of the lamellae maximum and the optic axes maximum in the undeformed fabric, to the elliptical cross section of the deformed cylinder proves the influence of intragranular slip in the deformation. The greater freedom of movement in the granular aggregate might be ascribed to the greater diversity of orientation of available glide planes in an aggregate as compared with a single crystal. Moreover it is possible that translation has operated to a considerable extent in the deformation of the marble whereas in the single crystal twinning was the most important mechanism of deformation.

It is difficult to evaluate the relative amount of translation-gliding and twin-gliding planes in the deformed fabric. In spite of the difficulty of measuring optic axes accurately in the relatively thick sections of deformed material it can be said with reasonable certainty that the majority of newly developed lamellae are the trace of *e* planes, which are known glide planes in calcite. But as calcite can glide on (*e*) by both translation and twinning this tells nothing about which process has operated. Where the individual bands of polysynthetic twins are broad enough to allow measurement of both optic axes the evidence of twinning is conclusive. Such measurable bands can rarely be found, however. Another clear evidence of twinning is the presence within the twin band of diagonal lamellae that make an angle with similar lamellae in the adjoining band. This also is rarely found. As pointed out by Buerger[43], a difference in extinction between two adjacent lamellae is not conclusive evidence of twinning because of the slight anisotropism likely to be caused by strain developed locally along glide planes. In many grains the color banding disappears when the lamellae are turned into a position where their bounding surfaces are parallel to the axis of the microscope. In thick sections this color-banding could be a wedge effect caused by the diagonal position of the boundary surfaces of discontinuity. When these surfaces are placed upright their traces become hair-sharp lines and in polarized light the whole surface of the grain shows a uniform interference color. So many of the grains that were examined lacked conclusive evidence for twinning that probably translation has operated to a marked extent in deformation.

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[43] Buerger, M. J. Translation-gliding in crystals. Am. Mineralogist, 15, 57, 1930.

As twin-gliding can only proceed in one direction sense, the possible field of operation by twin-gliding for a given field of lamellae maxima can be computed, where the preferred orientation of the optic axes is also known, by means of a cyclographic[44] projection. Such a computation was prepared by Dr. James F. Bell on a Leitz stereographic net and was kindly furnished to the author. Figures 45 and 46 show adaptations on an equiareal net[45] of Dr. Bell's projection of glide plane and twin in a calcite crystal. It is easy to see from figure 45 that any pressure applied *against* the direction *into which* the calcite twin axis would move from the original untwinned position would prevent twinning in the area stippled in the figure. Pressure applied in the unstippled area would allow twinning.

In figure 46 the heavily stippled area represents the field in which twinning is impossible on all the three potential glide planes. In the areas occupied by the open circles twin-gliding could operate on one plane. In fields indicated by the horizontal ruling, twinning could function on two planes and in white areas twinning could function on all three planes. Translation could function on any of these planes, as it can operate in both senses of a direction.

Figure 47 was constructed by rotating figure 46, 64° from right to left to bring the optic axis of the single crystal into the position of the optic axes maximum in the fabric of the undeformed Yule marble. This brings the poles of (e_2) and (e_3) into the position of those lamellae in which the resolved shear stress on the glide systems would be great enough to allow gliding. The heavily stippled areas in this figure again represent fields in which compressive force would prohibit twin-gliding on all (e) planes. Twinning could operate to a limited degree in the areas occupied by open circles and horizontal ruling, and could function on all potential glide planes in the white areas. Compressive force was applied to Yule marble in the positions shown by P, Q and R in the diagram.

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[44] A cyclographic projection is one in which the traces of planes are plotted as great circles.

[45] In this work the equiareal net has been used entirely in order to keep the results comparable although not entirely true in angular relation.

The factor that controls the initial movement in plastic flow is the position of potential glide systems in the granular aggregate with reference to the direction of applied force. In figure 20 the dashed lines indicate the border of the field that would contain potential *e* glide planes related to the axes maxima in the Yule fabric. Plastic flow in the Q and R cylinders could start on *e* planes occupying any maximum except those in the east-west position, in which position the resolved shear stress would be zero for applied force that is either parallel or normal to the east-west direction, i.e., for Q, R, or P cylinders. In the P cylinder plastic flow could not start upon glide lamellae in any peripheral maxima because they are parallel to the P direction. However it could start on potential glide systems occupying other positions within the field of *e* planes related to the optic axis maximum (See fig. 20).

A comparison of figure 20 with figure 47 indicates that when compressive force is applied in the Q direction the majority of glide lamellae are in the field where twinning would be a minimum. Pressure applied at R or at P, on the other hand, would be relatively favorable for twin-gliding.

The course of compressive deformation in R cylinders moves the lamellae maxima close to the R position in figure 47, i.e., closer to the field most favorable for twinning. In the P cylinder deformation moves the lamellae towards the position of P in figure 47. In this position the opportunity for twinning is much greater on one side of the east-west direction than it is on the other, a fact that may possibly be connected with the asymmetry of the lamellae shown in figures 37 and 40, which are diagrams of the P cylinder after deformation. The diagrams of the R cylinder after deformation are slightly asymmetrical but somewhat less so than the P cylinder diagrams. Further work on deformed cylinders will be necessary to establish the real significance of this asymmetry in the fabric.

Theoretically the operation of twin-gliding should be at a maximum in the deformation of the R and P cylinders. By the same reasoning compression of the Q cylinder would be least favorable for twin-gliding because the pressure is applied in a field unfavorable for twinning and because the end position of the glide lamellae would be in the heavily stippled area in figure 47 where twinning is impossible. In the P cylinder twin-gliding should operate to a more limited extent than in the R cylinder.

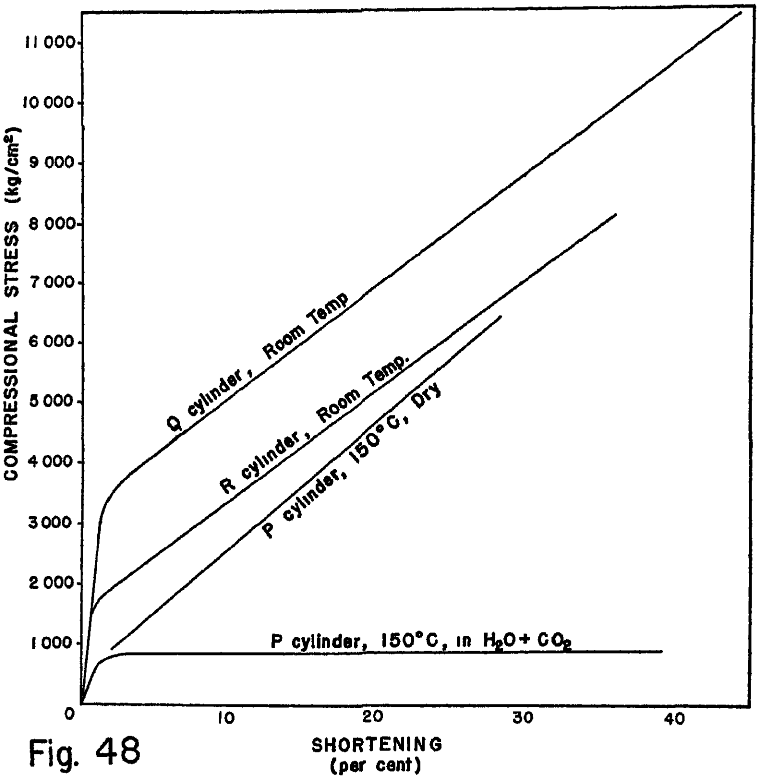

Fig. 48

Figure 48 Generalized stress-strain curves for Yule marble deformed in compression under 10,000 atmospheres confining pressure. Smoothed experimental data. (Courtesy of David T. Griggs.)

In this connection it is interesting to compare the strength of differently oriented cylinders. The difference in their reaction to compressive force is shown in figure 48, which shows the stress-strain curves for the deformation of P, Q, and R cylinders. This graph was kindly furnished by Professor Griggs, who emphasized the fact that the data from which he has constructed the curves, were unsatisfactory from the standpoint of physical measurement because they represent early experimental work before the apparatus and technique

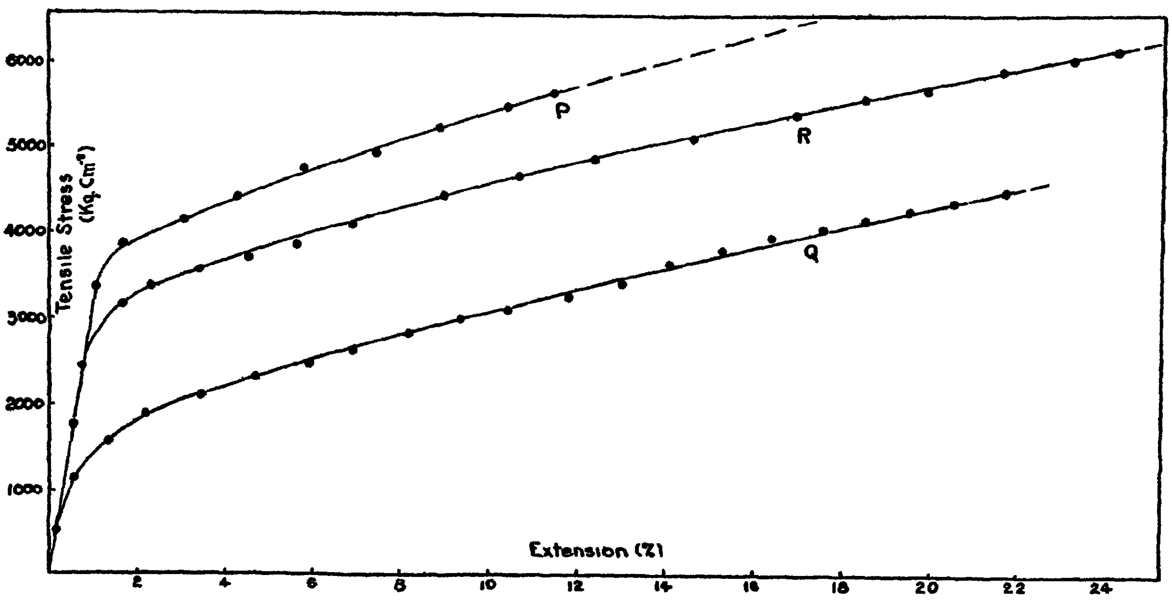

Figure 49 Stress-strain curves for Yule marble deformed in tension at 10,000 atmospheres confining pressure (Balsley, J R, Nat. Res Council, Am. Geophysical Union, Transactions for 1941, p 522.)

were perfected. He is hoping to run new compression tests with the most recent improvements in technique. The stress and strain determinations from which the generalized curves in figure 48 were plotted have been corrected as far as possible for change in shape of the cross section during compression, etc. and they probably represent what actually happened as well as can be reconstructed from the available data.

Comparison of these curves with Balsley's stress-strain curves in tension shown in figure 49 suggests a relation between the possibility of twin-gliding and strength in compression, similar to that suggested by Balsley[46] for deformation in tension. Positions favoring the operation of twin-gliding in tension are exactly reversed from the corresponding positions in compressive strain. Q is the cylinder that should expectably be the most favorably oriented for twin-gliding in tension and R and P should be the least favorable orientation for twinning, whereas in compression the Q cylinder is the least favorable and the R cylinder the most favorable for twinning.

In the following table strength of Yule marble in compression and in tension is compared.

	Deformation in compression				Deformation in tension		
	(measured in Kg/cm^2 at 10,000 atmospheres confining pressure)						
	Dry			in H_2O CO_2	Dry		
	20°C		150°C		20°C		
	Q	R	P	P	P	Q	R
Plastic yield point	3200	1500	~ 700	700	3400	~ 500	2500
10% strain	5100	3400	~2600	850	5400	3000	4600
20% strain	7000	5200	~4700	900	~6900	4200	5600
30% strain	8800	7000	~6800	900	...	.	...

From this table it is easy to see that strength in *compression* is a maximum in the Q cylinder where twin-gliding would be at a minimum and conversely it is a minimum in *tension* in the Q cylinder where twin-gliding could be a maximum. This suggests, as pointed out by Balsley, that twinning has played an important part in deformation. The low values for strength in compression in the two P cylinders as contrasted with the values for R and P cannot be compared because P was not deformed under conditions comparable to the deformation of Q and R. A rise in temperature definitely facilitates compressive strain in both dry and wet deformation. In the presence of carbonated water, where recrystallization takes place, deformation continues after the plastic yield point, even with no appreciable increase in differential stress

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[46] Balsley, Jr., J. R. Deformation of marble under tension at high pressure. Am. Geophys. Union Trans. Pt. II, 519-525, 1941.

So far it is impossible to make a reliable estimate of the relative importance of twin-gliding and translation-gliding. We know, as shown by Bell, that translation-gliding can function under conditions where twin-gliding is impossible, but the relative ease of compressive and tensile deformation in cylinders that are oriented so that twin-gliding is permissible suggests that translation is facilitated to a large but undeterminable extent by twin-gliding, which, wherever it can operate, adds to the ductility. The relative importance of these two mechanisms of slip cannot be determined from evidence at hand.

CONCLUSIONS

Theoretically the possibility of intragranular grain-gliding in a monomineral polycrystalline calcite aggregate would be governed by the law of maximum resolved shear stress. For the fabric of the Yule marble this would determine the position where gliding on the $\{01\bar{1}2\})$ glide system would commence as a result of compressive force applied parallel to the cylinder axis in the P, R and Q cylinders. In the R and Q cylinders all maxima shown in figure 20 except the east-west maximum would contain grains on whose glide systems the resolved shearing stress would be a maximum. The susceptibility to glide in the grains occupying the east-west maximum would be a minimum but here translation-gliding would gradually change the shape and position of the operating grains and could thereby bring glide lamellae from the position of the east-west maximum into a position with reference to the applied force where they could glide. In the P cylinder the resolved shearing stress on the peripheral maxima in figure 20 would be so small that the movement would start on grains outside these maxima. Gliding in grains within these maxima would then gradually come into play. The most that can be said as a

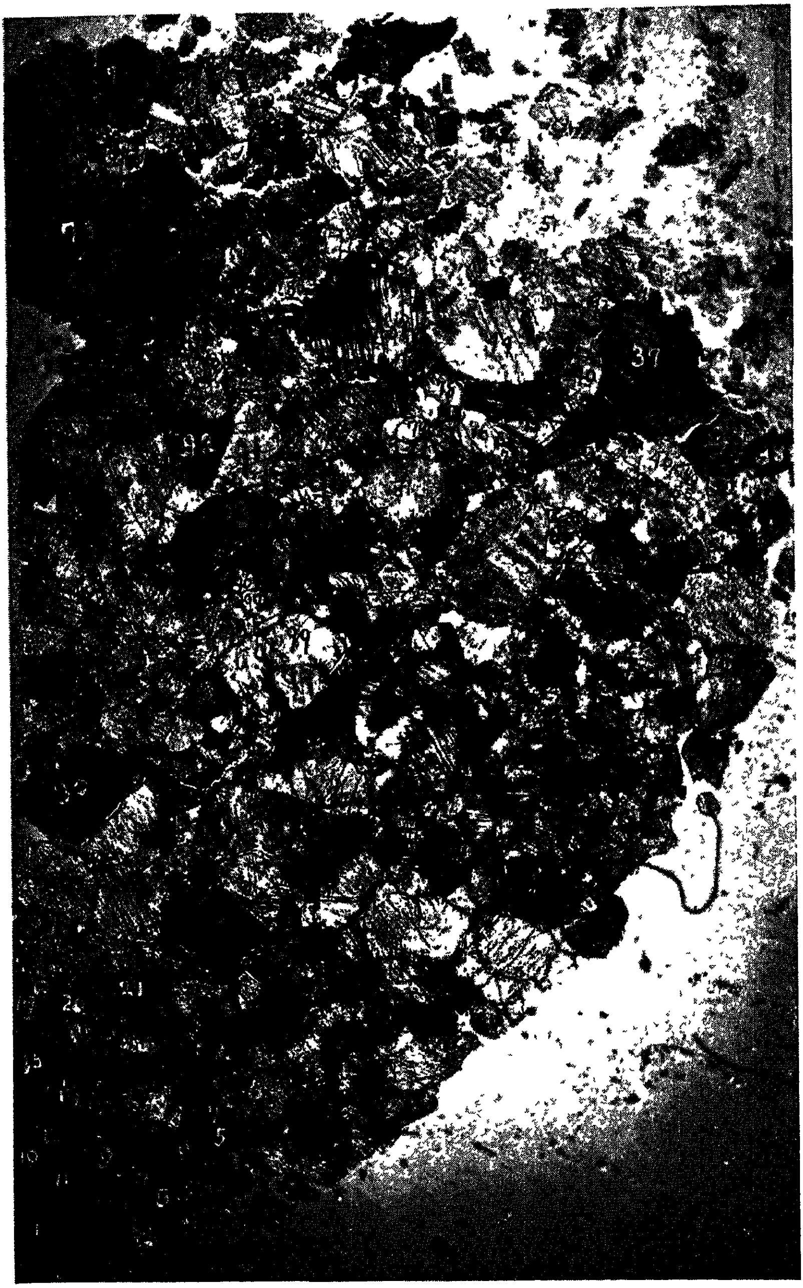


Plate 5 Thin section from R cylinder after deformation in compression (dry at 20°C) × ca 15. Section cut normal to the cylinder axis, which is the axis of compressive force The east-west axis of the elliptical cross section of the deformed cylinder is vertical in the photograph

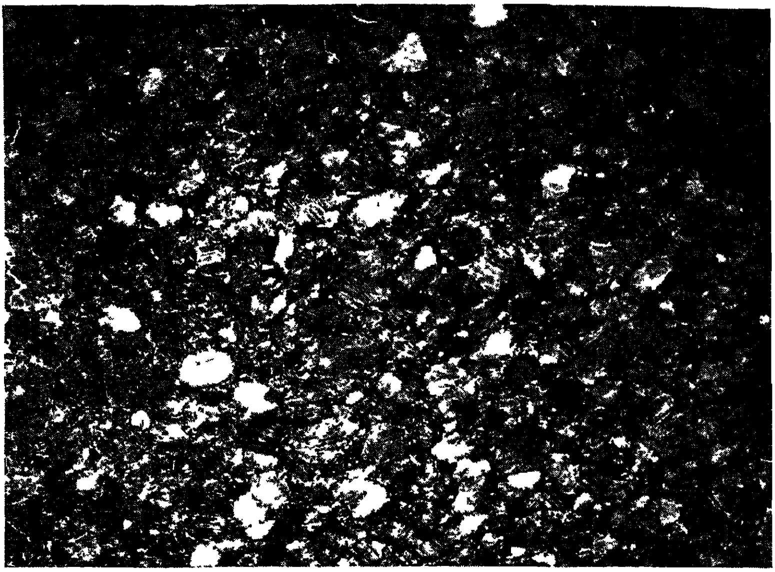


Plate 6 Thin section from R cylinder after compression (dry at 20°C) x ca 10 Section is parallel to P face of thin-section block, i e normal to north-south direction The axis of compression was the vertical direction in the photograph Note the elongation of the grains in a direction normal to the axis of compression.

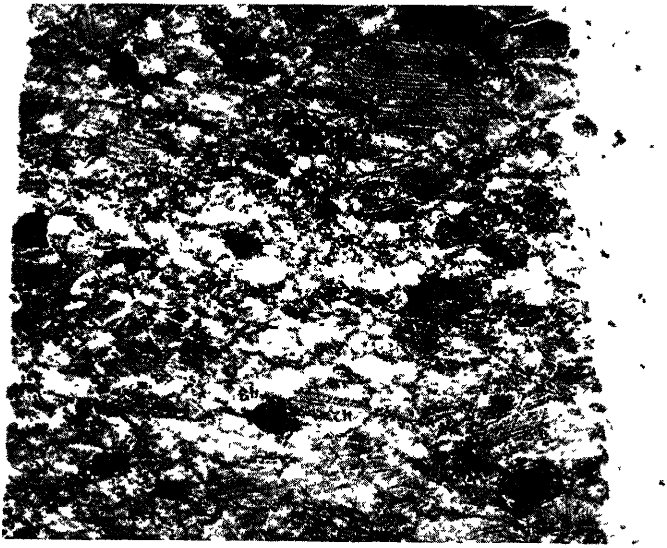


Plate 7 Thin section from R cylinder after compression (dry at 20°C) x ca 12 Section is parallel to Q face of thin-section block, i e. normal to east-west direction The axis of compression was the vertical direction in the photograph Note the elongation of the grains in the direction normal to the axis of compression

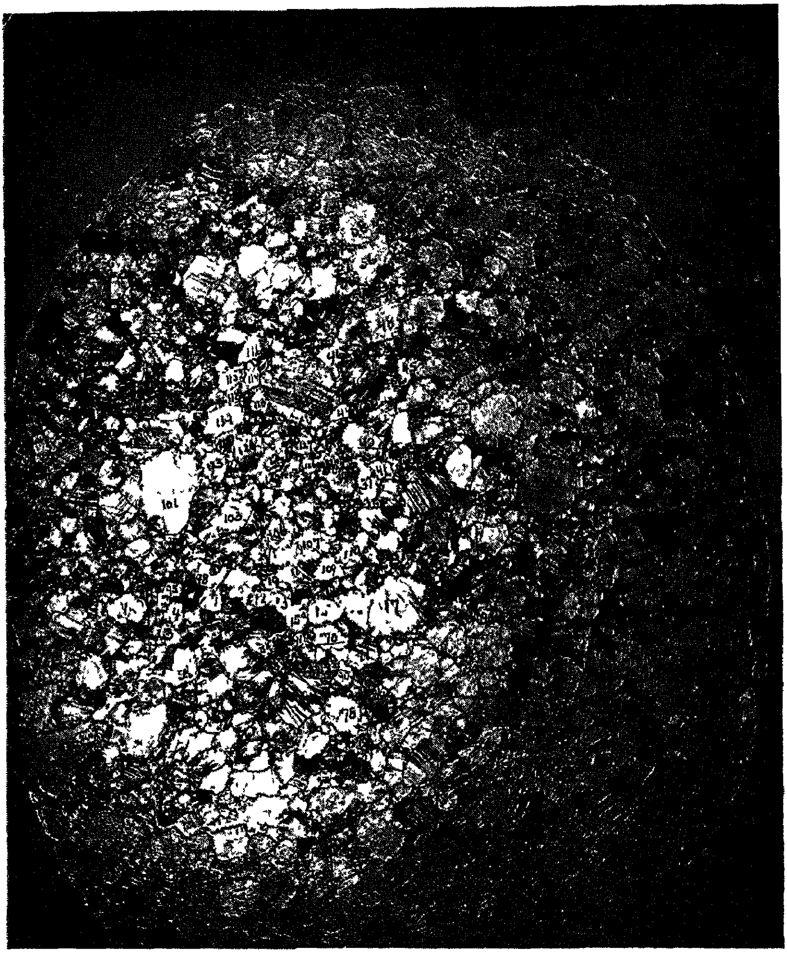


Plate 8 Thin secton from P cylinder after compression (dry at 150°C) x ca 15 Section is normal to the cylinder axis, which is the axis of compression The east-west direction which is the major axis of elliptical cross section is vertical in the photograph

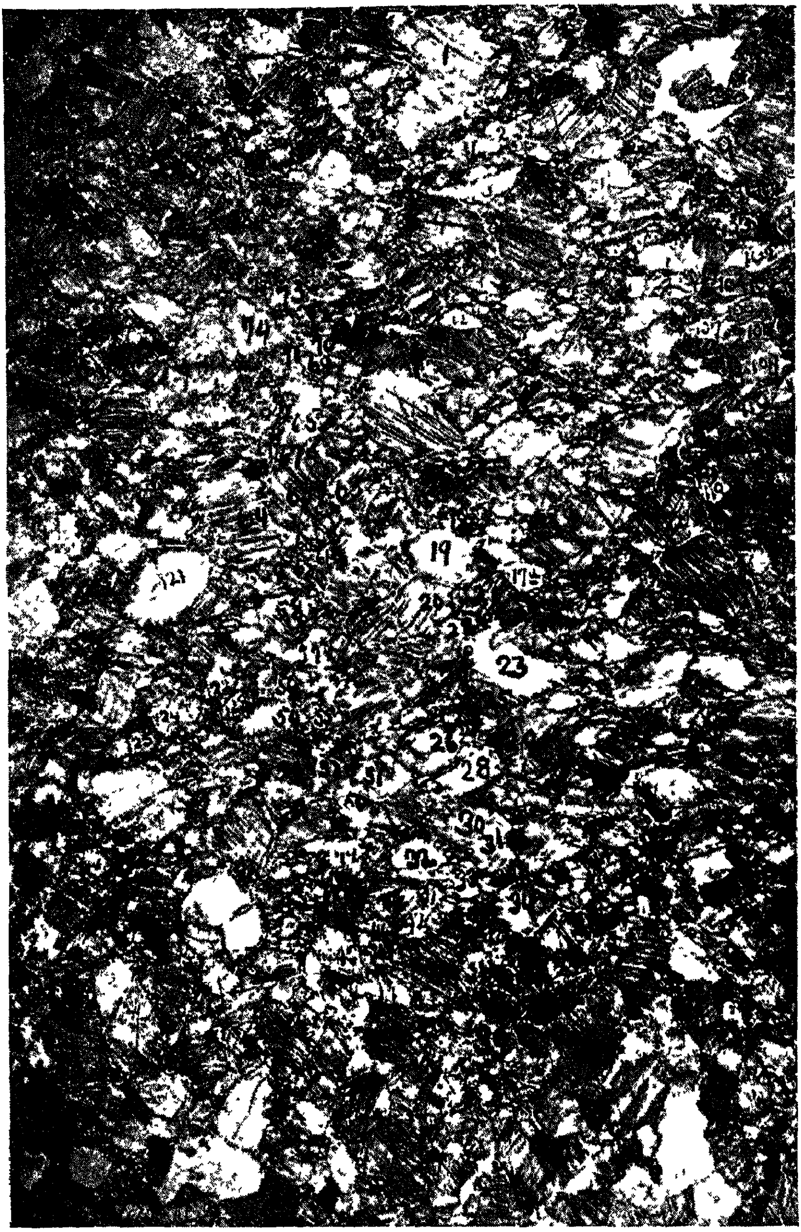


Plate 9 Thin section from P cylinder after compression (dry at 150°C) x ca 15. Section is parallel to R face of thin-section block, i e normal to the vertical direction Axis of compression was parallel to the vertical direction in the photograph Note the elongation of the grains in the direction normal to the axis of compression

Plate 10 Thin section from P cylinder after compression (dry at 150°C) × ca 20 Section is parallel to Q face of thin-section block, i e normal to the east-west direction Axis of compression was parallel to the vertical direction in the photograph Note the elongation of the grains in the direction normal to the axis of compression.

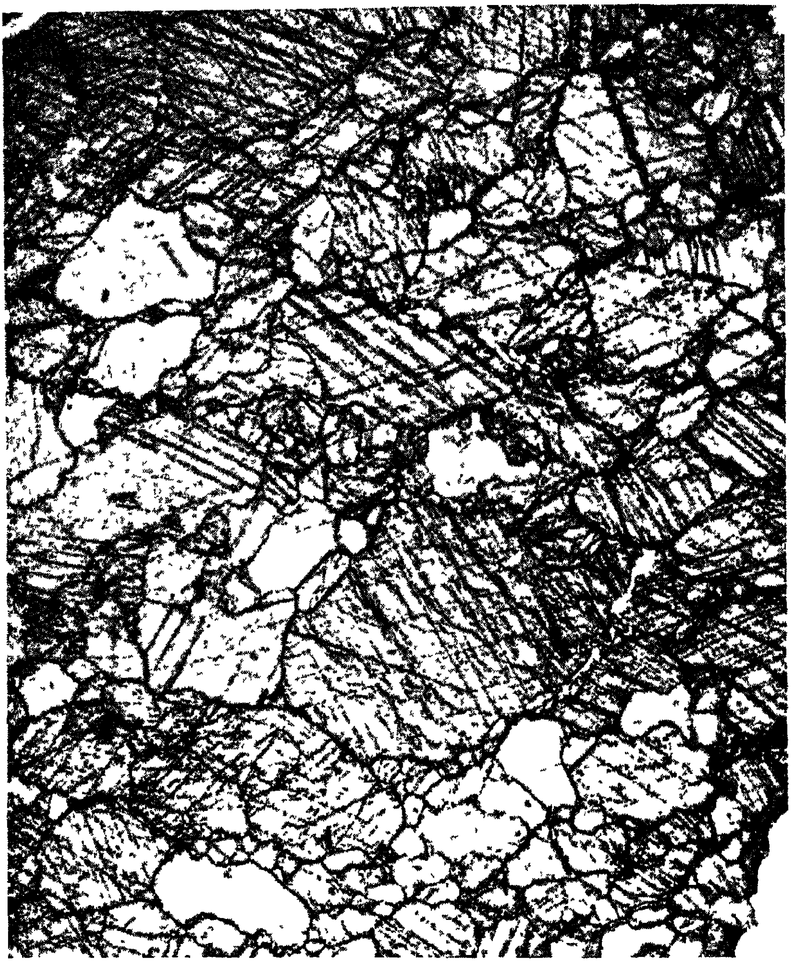


Plate 11 Thin section from P cylinder deformed in compression at 150°C in carbonated water x ca 45 Section is parallel to Q face of thin section block Note the recrystallization fabric

result of the present study of the deformed fabric is that the movement picture, indicated by the diagrams, shows clearly a rearrangement of the planar grain structure (*ab*) into a planar structure (a_2b_2) that is approximately normal to the compressive (see fig. 50). The twofold maximum that appears in the lamellae measurements in diagrams from both the R and the P cylinder is strongly suggestive of intragranular slip on two intersecting sets of slip planes, which is the "plättung" or flattening slip developed by the operation of two intersecting slip planes within the individual grains. This has been described by Sander[47] as a possible mechanism by which a foliation normal to the applied force is developed. In order to confirm this slip mechanism it will be necessary to measure, grain by grain, optic axes and lamellae so as to plot the crystallographic position of the lamellae and the positions of possible glide directions $[01\bar{1}2 : 0001]$, $[01\bar{1}1]$, $[01\bar{1}2]$. This will require thinner sections than were cut for the preliminary deformation work and such sections can doubtless be cut in future work.

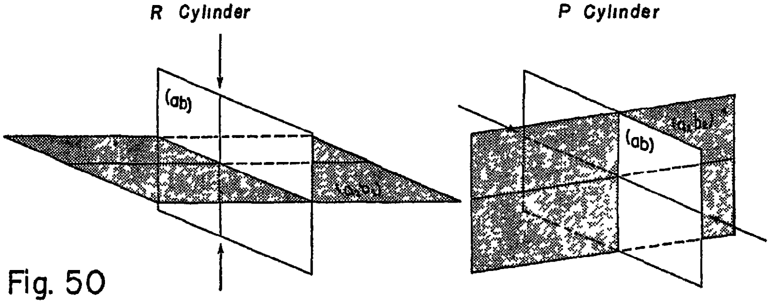


Fig. 50

Figure 50 Diagram showing rearrangement of planar structure in Yule marble as a result of deformation in compression. Arrows indicate direction of applied force. Unshaded planes show position of slip plane. (*ab*) in the unstrained fabric. Shaded planes show new positions of slip planes (a_2b_2) after strain.

The present study of fabric before and after deformation has established that a radical change in preferred orientation, both dimensional and crystallographic, has resulted from the deforming movement and that this change is definitely related to the fabric pattern of the undeformed marble. Beyond this, its chief value lies in pointing the way to individual problems of the deformation and to the way in which they may be attacked in future investigations.

[47] Sander, Bruno. Gefügekunde der Gesteine, 219-220, Wien, 1930

COMPRESSION TESTS ON ORIENTED SPECIMENS OF YULE MARBLE

HENRY A. LEPPER, JR.

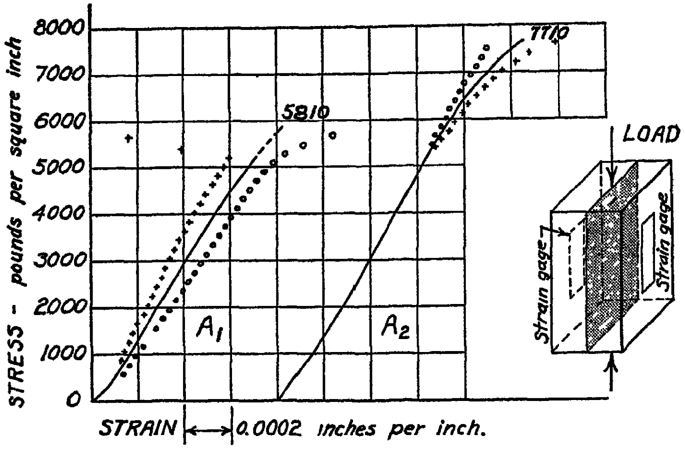
IN connection with her extensive studies of Yule marble, Mrs. Eleanora Bliss Knopf requested the writer to make compression strength tests on oriented samples of this material. Four test samples were prepared by Mrs. Knopf and were tested in the Materials Testing Laboratory, Department of Civil Engineering, Yale University, during the latter part of July, 1948.

Two pairs of specimens having known orientation with respect to the planar "grain" of the marble were prepared. The "grain" of the marble, recognized by the quarryman as the plane of easiest splitting, is a structure, determined by the dimensional arrangement of ellipsoidal calcite grains, oriented so that the shortest axis of the ellipsoid is normal to the grain. This dimensional arrangement is accompanied by a preferred orientation of the crystal lattice, determined by a statistical concentration of the calcite optic axes in a position that makes an angle of 26° with the "grain" structure.

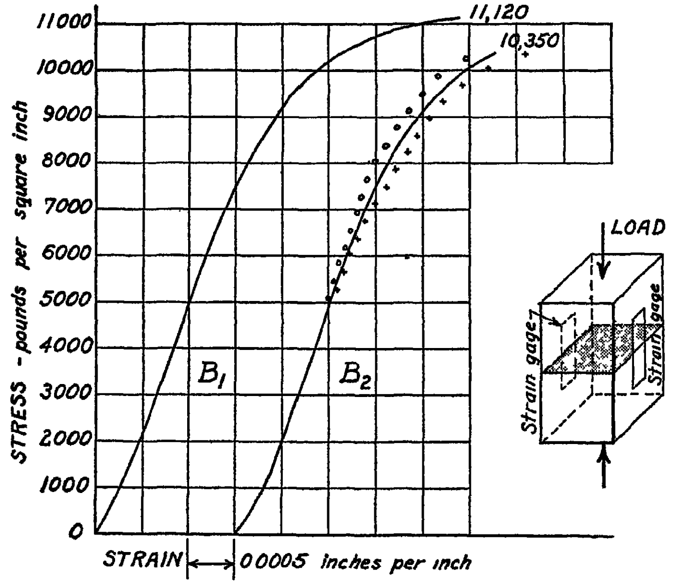
The test specimens were prismatic, nominally one inch square cross-section and two inches long, with smooth finish on all surfaces. Specimens "A" were cut so that the compression force was applied parallel to the planar grain; in the other specimens "B" the force was applied normal to the grain. The direction of loading with respect to the grain and the location of the gages used to measure the strain in the direction of loading are shown on the sketches in figure 2.

Apparatus: A Southwark-Tate-Emery Universal Hydraulic Testing Machine of 120,000 lbs. capacity was used for the tests. The 12,000 lb. loading range of the machine was adequate for these specimens. This range could be read to 10 lbs. and has been calibrated to show an error of less than ± ½ per cent. In such a machine the specimen is subjected to a slowly increasing, steady strain while the load transmitted through the specimen as a result of the strain is indicated on the dial of the machine.

Strains were measured by "SR-4" resistance wire strain gages connected to a Baldwin-Southwark Portable Strain Indicator, Type K. These gages consist of a grid of fine wire



A. LOAD PARALLEL TO "GRAIN"



B. LOAD NORMAL TO "GRAIN"

FIGURE 2. STRESS-STRAIN DIAGRAMS FOR YULE MARBLE TESTED IN COMPRESSION.

cemented to a paper back which in turn is cemented to the specimen. The cement bond is adequate to cause the wire to strain with the specimen, which strain produces a change in electrical resistance of the gage element. The strain indicator is essentially a bridge circuit for measuring this change in resistance and consequently the strain in the specimen.

Procedure: Each specimen was measured and strain gages were cemented in place at the centers of two opposite sides. Another gage to provide a balancing resistor for the indicator and temperature compensation in the strain readings was cemented to another piece of the marble which was placed on the testing machine near the test specimen.

Each specimen was mounted in the testing machine between solid bearing plates by seating the bearing faces in wet plaster of paris and then running the heads of the machine together until the excess plaster was squeezed out, leaving a thickness of 1/32 inch or less at each end. No appreciable load was required to do this. The plaster was allowed to harden overnight so that the specimen was firmly bedded in place for testing. This method of mounting the specimen was chosen to assure throughout the sample, uniform strain, which, if the material is statistically homogeneous, will provide uniform stress distribution on the cross-section. Better results are obtained this way than by the more commonly used spherical bearing block, with its attendant difficulties in accurate centering. A close-up view of the specimen placed in the machine is shown in figure 1. The felt pad covering one strain gage is seen on the side of the specimen, with the block for the compensating gage beside the test sample.

Loading during the test was continuous and at a practically constant rate of strain. Strain readings were taken at frequent load increments, each gage being read at alternate load points. Thus the loading could be kept steady and the two gages read and recorded. Maximum load was noted but because of the sudden failure it was not possible to measure the strains at this point. Following the test, each specimen was carefully removed from the machine to permit examination of the fracture.

Results Complete stress-strain diagrams for the four tests are shown in figure 2. The plotted curves are based on the average of the two strain gage readings, with individual gage readings shown where there is appreciable divergence from the average curve.

The spread between the strains read on the two gages is a measure of the axiality of the loading. Agreement is quite good in all cases except specimen A_1, the first tested. In this case, the plaster used for bedding was probably not thoroughly hard, and the strains show appreciable differences. For specimen A_2 the strains agree well to a stress of about 6,000 lbs. per sq. inch, then show a divergence. This is believed to be due to nonhomogeneity in the material. Specimen B_1 shows excellent agreement to failure; B_2 is not so good but still satisfactory.

Strength results from these curves are summarized in Table I. Compressive strength is greater for loading normal to the plane of the "grain," but the stiffness of the marble (modulus of elasticity) is less than for loading parallel to "grain." The marble is capable of deforming about three times as far when loaded normal as when loaded parallel to grain. Thus the strength properties reflect the aelotropic nature of the marble.

TABLE I

Results of Compressive Tests

Specimen	Orientation	Compressive Strength lbs./in ^{2}	Modulus of Elasticity lbs /in.^{2}	Maximum Strain in./in.
A_1	Load parallel to grain	5,810	8,340,000	0008
A_2	" " " "	7,710	9,000,000	.0010
B_1	Load normal to grain	11,120	5,420,000	.0038
B_2	" " " "	10,350	5,850,000	0028

For a relatively brittle material, such as this marble, it is essential to secure uniform stress on the test specimen to give a correct value of compressive strength. If the test loading is eccentric, bending stresses will be introduced which produce a maximum stress appreciably greater than the average value calculated as the load divided by the cross-sectional area of the specimen. Failure occurs when this maximum stress reaches the strength of the material regardless of the value of average stress at the time. Such failure is progressive, starting at the high-stress region, and when a portion of the area has failed, the load must be supported by the remainder of the area, thus completing the failure. The maximum load then will be less than that required to bring the entire cross-section up to the compressive strength. The average stress computed for this maximum load and reported as the ultimate strength will be appreciably less than the true ultimate strength.

Comparison of specimens A_1 and A_2 illustrate this point. A_2, with better loading conditions, showed 7,710 lbs. per sq. in. ultimate strength while A_1 gave only 5,810 lbs. per sq. in. The value for A_2 is felt to be a better result than A_1. The "B" specimens are in reasonably close agreement.

Specimen A_1 apparently had some kind of flaw in it which may have induced a premature failure. Evidence of this is seen in the readings of gage 2, which showed a decrease in strain after about 5,300 lbs. per sq. in. This was probably caused by a partial failure of the material which relieved a portion of the specimen of its stress. However, no further evidence of a flaw could be found upon visual examination. No unusual behavior was noted in any of the other tests.

Compressive strength of a material of this sort is affected by the proportions of the specimens and the end bearing conditions. The fully restrained ends of these tests will give higher strength results than if the specimens were of the same proportions but not so restrained at the bearing faces. Cube specimens would probably give higher strength, while a larger length-width ratio would give somewhat lower strength than these tests. Cylindrical specimens having a length-width ratio of two-to-one would probably show somewhat higher strength. In concrete, which is in some respects similar in its compressive properties to this marble, these effects have been studied in some detail.[1] These factors connected with the test itself illustrate the difficulty of determining the compressive strength of such material in a confined state or in large masses. Tests of the sort made here are chiefly useful for showing relative strength values and not absolute strength of a massive material.

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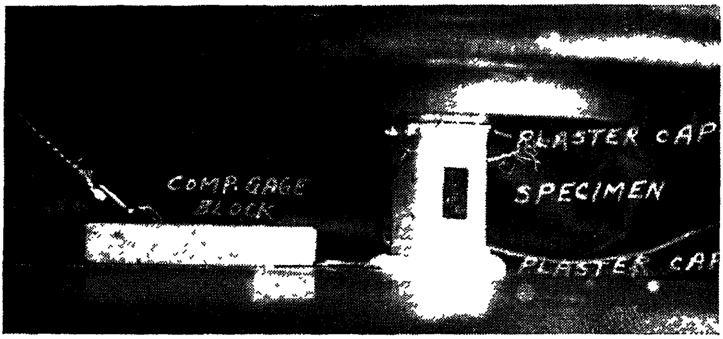

Figure 1 Close View of Specimen in Position for Test

QUANTITATIVE AREAL GEOLOGY OF THE UNITED STATES

DONALD V. HIGGS

ABSTRACT. The areal exposure of each of the principal systems and series discriminated on the Geological Map of the United States (U. S. Geological Survey, 1933) has been determined from that map. Utilizing these data and the estimates of Holmes (1947) as to the durations of the corresponding periods and epochs, it is apparent that for each million years of geologic time the areal exposure of the corresponding rocks tends to diminish sharply with geologic age. This relation indicates that the probability of exposure of evidence for equivalent historical events likewise diminishes with geologic age. We can arrive at a crude quantitative estimate of the loss of record with time—a measure of "geological perspective."

INTRODUCTION

IN 1928 David White introduced a spiral graph "illustrating geologic time and the evolution of the earth's history." The object was to "present an appropriate, stimulating, and more adequate form of graph for use in developing the mental picture of the lapse of geologic time and superposed geologic history." The age determinations were approximately in accordance with the calculations of Holmes, Lawson, Schuchert and Barrell (White, 1928). More information is available now and, although the spiral graph is a striking representation, a somewhat different approach using more recent data may be worthwhile. This paper is based on quantitative data from the United States. By comparing areal exposure with duration of corresponding time units (as established by Holmes, 1947) we can estimate the probable completeness of the geologic record, and hence our chance of observing details of equivalent events in the different periods of geologic history.

Dr. James Gilluly, Professor of Geology at the University of California at Los Angeles, suggested the problem. His advice and criticisms are greatly appreciated.

PROCEDURE AND RESULTS

The area of exposure of each of the major units recognized on the Geological Map of the United States (U. S. Geological Survey, 1933) was determined by cutting up a map of the United States and weighing, separately, the paper pieces corresponding. The percentage of the total area of exposure within the continental limits of the United States then was computed for each of the units. The results are given in table 1. Data for the major units (table 2) were obtained by combining the values of the smaller subdivisions. In the cases of undifferentiated units such as Quaternary-Pliocene sediments and extrusives, Pliocene-Miocene sediments and extrusives, et cetera, the values were divided equally among the periods or epochs concerned.

The estimates of areal exposure for the major divisions of geologic time are listed in table 2. These values were plotted on a graph, figure 1, to show the relative magnitudes of exposure. The values then were divided by the corresponding duration of time and the results (square miles/millions of years) were plotted as a factor graph, figure 2. Figure 3 is an isometric diagram representing the area of exposure plotted against time.

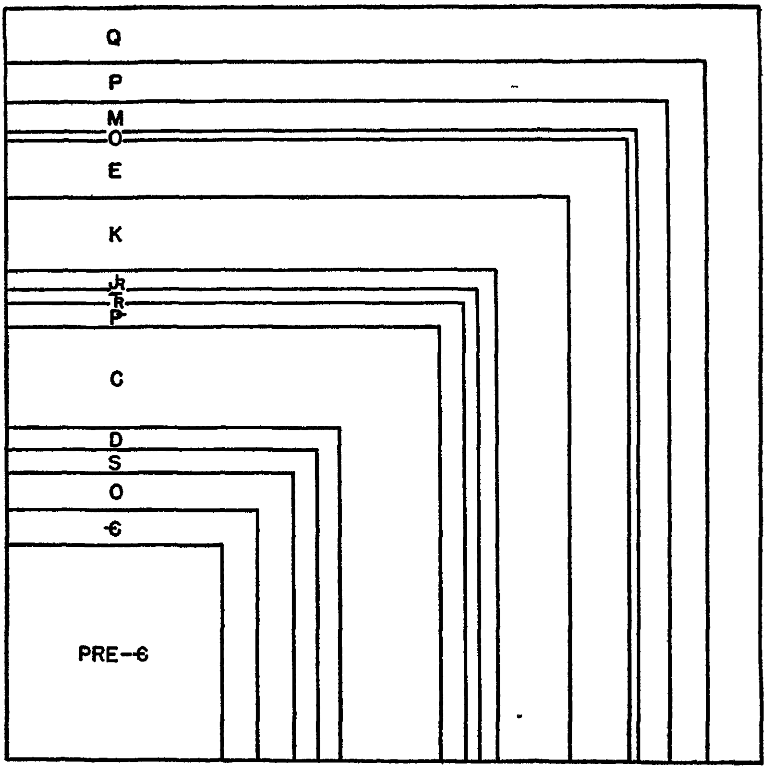

Figure 1. Cumulative graph representing areas of exposure of major rock units shown on Geologic Map of United States (column 1, table 2).

TABLE 1

Units Considered	Wgt. (gms.)	percent
Quaternary sediments	16.4682	12.43
Quaternary extrusives	.099	.074
Quaternary-Pliocene sediments	.580	.437
Quaternary-Pliocene extrusives	3.929	2.965
Pliocene sediments	3.985	3.007
Pliocene-Miocene sediments	5.619	4.241
Pliocene-Miocene extrusives	2.607	1.967
Miocene sediments	5.154	3.890
Oligocene sediments	1.128	851
Eocene sediments	13.047	9.847
Eocene extrusives	.978	.738
Tertiary extrusives	3.477	2.624
Tertiary intrusives	.372	.280
U Cretaceous sediments	14.853	11.232
L Cretaceous sediments	506	.381
Cretaceous sediments	3.058	2.326
Cretaceous-Jurassic sediments	.708	.534
Jurassic sediments	.803	.606
Jurassic intrusives	2.748	2.074
Jurassic-Triassic sediments	1.007	760
Triassic sediments	1.673	1.262
Permian sediments	4.891	3.691
Pennsylvanian sediments	10.392	7.842
Mississippian sediments	5.643	4.250
Carboniferous sediments	1.075	.811
Carboniferous extrusives	.301	.227
Devonian sediments	3.353	2.530
Silurian sediments	3.003	2.266
Silurian-Ordovician sediments	539	406
Ordovician sediments	3.828	2.889
Ordovician rocks	.198	.149
Ordovician-Cambrian sediments	305	.230
Cambrian sediments	2.385	1.800
Cambrian intrusives	846	.638
Cambrian rocks	.236	.178
D-S-O-C sediments	.378	.285
Paleozoic undifferentiated	.429	.323
Paleozoic extrusives	049	.036
Paleozoic intrusives	.602	.454
Pre-Cambrian sediments	5.491	4.144
Pre-Cambrian extrusives	.714	.538
Pre-Cambrian schist and ign injec.	536	404
Pre-Cambrian sed and ext.	848	.640
ph — age unknown	.094	.070
Pre-Eocene sed. and ign. complex	.214	161
Pre-Franciscan intrusive	053	.040
	132.485	99.976

TABLE 2

	Area of exposure[1] in sq miles	Percent of total exposure	Duration[2] (millions of years)	Area/millions of years
Quaternary	421,600	14 17	1	.421
Pliocene	254,000	8.54	11	.021
Miocene	229,600	7.72	14	.0164
Oligocene	47,000	1 58	12	.003917
Eocene	336,300	11.31	20	.01683
Cretaceous	442,600	14.21	69	.006120
Jurassic	99,080	3.33	25	.003961
Triassic	48,800	1.64	30	.001626
Permian	114,000	3.83	21	.005420
Carboniferous	395,300	13.28	52	00760
Devonian	81,530	2 74	58	.001406
Silurian	79,800	2.68	37	.002150
Ordovician	106,300	3 57	80	001328
Cambrian	87,510	2.94	80	.001094
Pre-Cambrian	250,800	8.43	2000±	.000125
TOTALS	2,974,220	99.97%		509977

[1] Douglas, 1930.
[2] Holmes, 1947.

There are many errors inherent in this procedure, but in part they should be compensating. Errors involved in cutting the map are apparently no greater than the inaccuracies of the map. Variations in the weight of the paper are unavoidable and they are considered to be evenly distributed throughout the periods. The possibility that most of the heavier parts of the paper, for example, are included in the area of the Cretaceous exposures would be extremely fortuitous. Errors undoubtedly result from the arbitrary assignments made in those cases where rocks of several periods or epochs are not differentiated. Where this occurred the values were divided equally among the units involved. The values for the Tertiary igneous rocks, for example, were divided equally among the four epochs.

CONCLUSION

The factor graph (figure 2) depicts a measure of the relative clarity of the geologic record. The relative degrees of accuracy possible in detailed analysis of particular events occurring in different periods of geologic time can be determined. Thus the probability of correct interpretation of the Pleistocene glaciation is about 80 times greater than for Permian glaciation and about 3,500 times greater than for Pre-

Cambrian glaciation. Similar relations exist for other problems where specific processes have been noted as occurring in different periods of geologic time. This includes mountain building processes, igneous activities, geosynclinal sedimentation, conditions of aridity and many others.

Figure 3 illustrates the size of the source area from which the history of a given interval of time is to be derived. In the

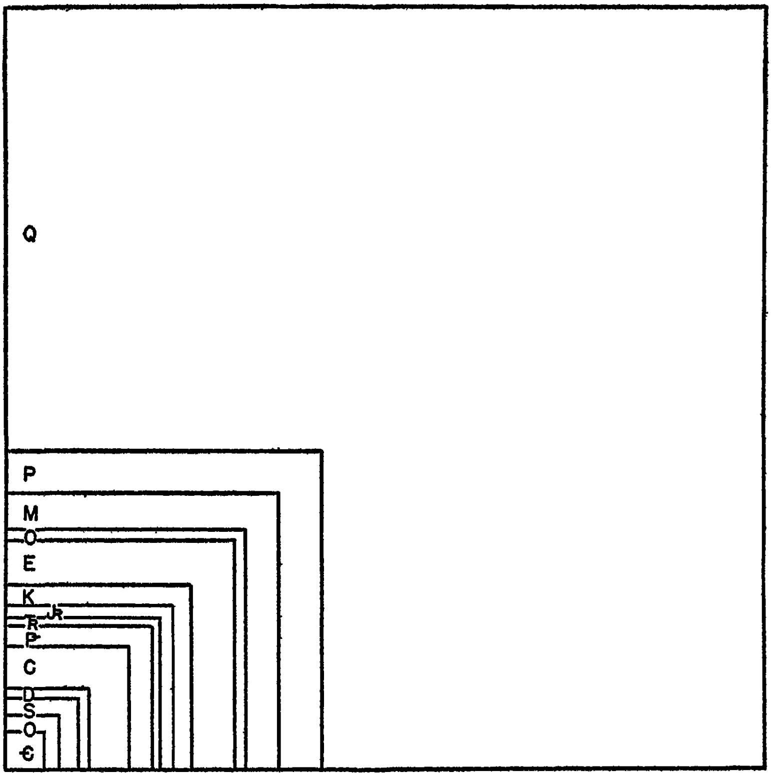

Figure 2 Cumulative factor graph showing relative clarity of the main units of the geologic record in area of exposure per million years of duration (column 4, table 2).

United States more than 75% of the earth's history (Pre-Cambrian) must be interpreted from less than 10% of the total area of exposure. Conversely about .05% of the record (Quaternary) is derived from 14.17% of the total area. Furthermore, rocks are altered by various processes after their formation and the older rocks generally are more changed. Therefore, the ratio of the areas of exposure presents the maximum relative clarity and not the complete record. Hence, increasing clarity is directly proportional to decreasing time.

Since the geologic record rapidly becomes more obscure as the time factor is increased, it is impossible to determine the relative importance of particular events occurring at different periods in the earth's history. It would be hazardous to state, for example, that orogenic movements during Pre-Cambrian time were of greater or less importance than orogenic movements during Tertiary time. A remarkably close correlation exists between curves representing thickness/time values and areal exposure/time values. The graphs, figures 4 and 5, were prepared by James Gilluly (1949), who very kindly permitted

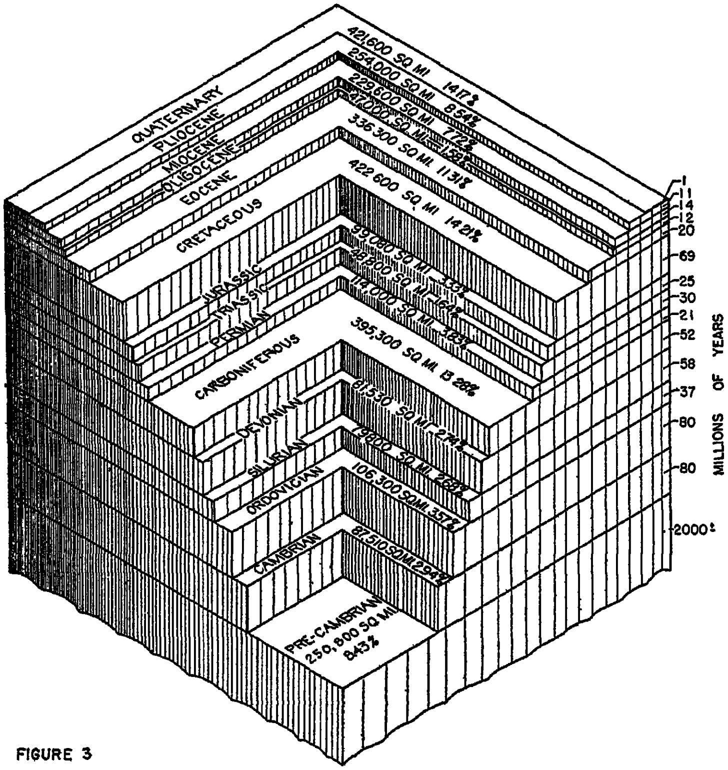

FIGURE 3
ISOMETRIC DIAGRAM WITH THE AREA OF EXPOSURE PLOTTED AGAINST TIME

Figure 3 Isometric diagram with the area of exposure plotted against time (columns 1 and 2, table 2).

their reproduction here. Figure 4 represents thickness per million years of the rocks corresponding to periods and epochs plotted against time since the middle of the period. The area of exposure for each million years of period length plotted against time since the middle of the period is shown in figure 5. Insufficient data prevent an accurate evaluation, but the trend of the curves is significant—constancy is the important implication. Deviations from the curve in figure 5 appear to be anomalous to the thesis. They may, however, be explained in part. Paleontologists are not in complete agreement as to the status of Oligocene. If it is grouped with the Eocene under the older term 'Eogene' epoch, the resulting value conforms to the general trend of the curve. Triassic and Jurassic

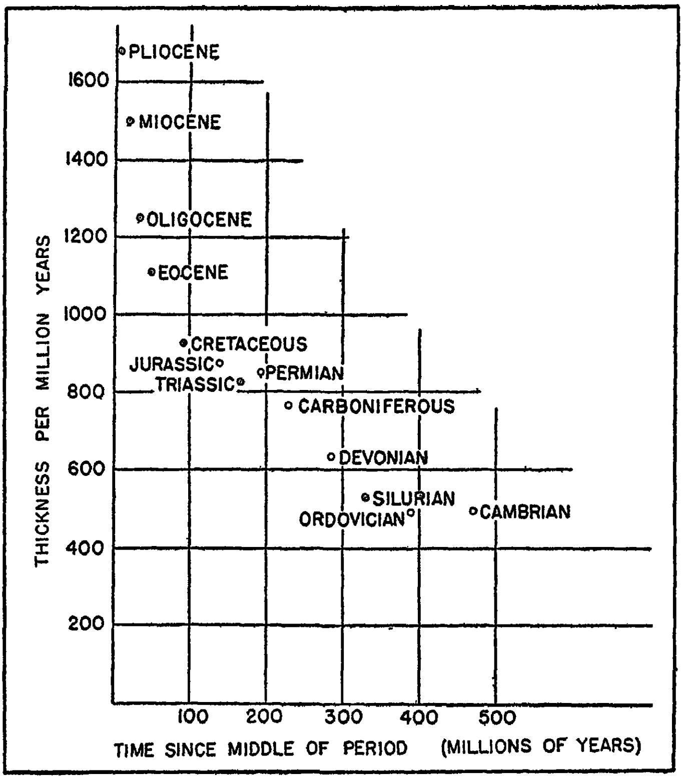

Figure 4 Thickness of rocks representing a period per million years plotted against time since middle of period (in millions of years). (Prepared by James Gilluly.)

deflections probably would have been less if a larger, and more representative, unit of the earth's surface had been used.

The information presented is applicable, in the strictest sense, only to problems within the continental limits of the United States. Further work is planned to secure data from

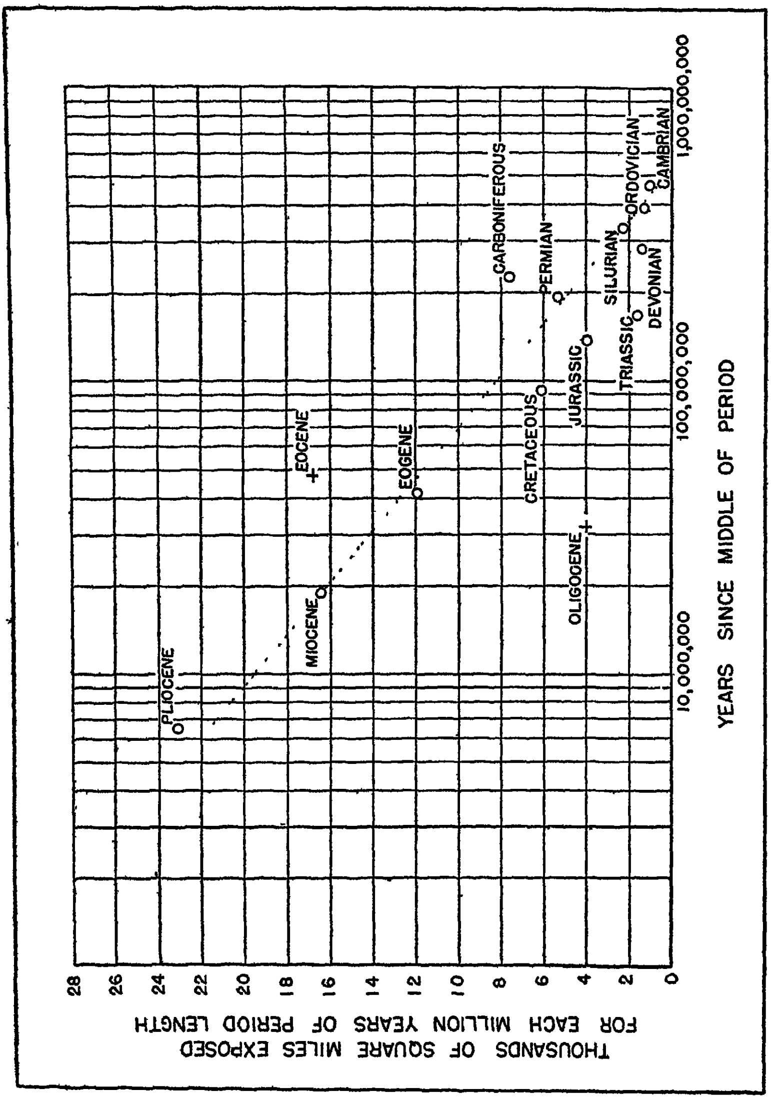

Figure 5. Thousands of square miles of exposure per million years of period length plotted against time since middle of period. (Prepared by James Gilluly)

considerably larger portions of continental exposures. The relative values for such areas probably will be little changed, except that a curve corresponding to that of figure 5 may exhibit deflections of smaller magnitude. Perhaps the present time is not unique.

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REVIEWS

The Atmospheres of the Earth and Planets; edited by GERARD P. KUIPER. Pp. vii, 366; 16 pls., 91 figs. Chicago, 1949 (University of Chicago Press, $7.50).—Telescopic observation has long proved inadequate to satisfy the general curiosity about the nature of the planets. Yet the physical analysis of their constitution has lagged behind the development of other branches of astronomy. As late as 1937 the Russian astrophysicist, B. P. Gerasimovič, the director of the great Poulkovo Observatory and former member of the Harvard faculty, lamented the neglect into which the study of the planets had fallen, while the exploration of the sidereal universe progressed rapidly. In particular, he urged the appreciation of the well-ripened methods of theoretical astrophysics to the analysis of planetary constitution. His own contribution, his last communication to reach the scientific world, was entitled "On the illumination of a planet covered with a thick atmosphere." The approximate solution given by Gerasimovič is superseded by recent work: Chandrasekhar has devised powerful mathematical methods capable of solving rigorously related problems of far greater complexity that only a few years ago would have seemed beyond hope of attack, like the distribution of polarization over the sunlit sky. Progress along the very lines Gerasimovič hoped for is admirably documented in the monograph under review. The editor is to be congratulated upon the organization of this symposium bringing together representatives of sciences usually regarded as far apart. Among the eighteen contributors are meteorologists, physicists, chemists, geologists and astronomers. Only some of the astronomical aspects will be commented upon here.

H. C. van de Hulst discusses "The scattering in the atmospheres of the earth and the planets," giving a comprehensive review of the physics of light scattering by particles of all sizes, based in part upon his recent Utrecht Thesis, which may be difficult to obtain in some places. This paper also contains a welcome summary of Chandrasekhar's solutions of problems of radiative transfer; Chandrasekhar's original publications are highly technical and treat only briefly of the physical background. Lyman Spitzer, Jr., deals with the rate of escape of gases from the gravitational field of a planet. Previous estimates of the time of escape must be increased by factors as large as 10^6 to allow for the fact that probably all the atmospheres are isothermal only at very great heights, with much lower temperatures in the stratospheres. Thus a temperature of 1500° K is required to disperse helium from the earth as fast as it is being liberated by denudation of rocks. Harrison Brown and the late Professor Rollin T. Chamberlin concur in the thesis that the earth's atmosphere is almost entirely of secondary origin, and that it was evolved as the result of chemical processes that took place subsequent to the formation of the planet. This conclusion makes it the more important to establish spectroscopically the retention of hydrogen by the giant planets. A direct spectroscopic test is suggested by G. Herzberg's discovery in the laboratory of quadrupole lines of H_2 in the photographic infra-red; spectrograms of these lines are reproduced. With the apparatus here described absorption spectra over optical path-lengths up to 5500 m can be obtained. Th. Dunham, Jr., describes the spectographic studies of the planets carried out at Mt. Wilson over the last twenty years, which culminated in the discovery of CO_2 on Venus and the definitive identification of NH_3 and CH_4 in the spectra of the giant planets. During the past three years G. P. Kuiper has observed the infra-red spectra with a photoelectric cell of the Cashman type and achieved such spectacular results as the proof that Mars' atmosphere contains a small amount of CO_2, and that the reflection spectra of Saturn's rings and of Mars' polar caps closely resemble that of frost or ice. Kuiper's contribution also contains numerous quantitative data on the chemical composition of planetary atmospheres (partly derived in collaboration with Herzberg) and an approximate theory of the chemical processes by which the present-day composition of the atmospheres was brought about.

RUPERT WILDT

The Face of the Moon; by RALPH B. BALDWIN. Pp. xiv, 239; 17 pls., 26 figs. Chicago, 1949 (University of Chicago Press, $5.00).—The author of this lively book is an astronomer supporting by new evidence the thesis that impacts of meteorites were responsible for the remarkable surface features of the moon. He points out that a simple quadratic expression suffices to represent the relation between depth and diameter of such diverse objects as shell craters, bomb craters, explosion pits, terrestrial meteorite craters, and lunar craters. The good fit of this equation is indeed remarkable, considering a variation in size by a factor of 10^5 of the objects studied. There is an interesting survey of the problem of fossil terrestrial meteorite craters and a thorough discussion of the chance of observing telescopically the flash of light of a meteorite colliding with the moon, which the author deems remote. Most of the book deals with matters under jurisdiction of the geologist rather than of the astronomer. The astronomical data the author brings to bear upon the early history of the moon (chapter 10) are, perhaps, not quite as reliable as the author believes, and his conclusions as regards the time scale of lunar evolution should be accepted with reserve. The extensive bibliography does not contain any references to modern attempts at explaining the lunar formations as result of orogenetic processes of the terrestrial kind. Conspicuous by absence also from other recent lists of reference is a memoir by Ingolf Ruud (Zeitschrift fur Astrophysik 8, 295-343, 1934).

RUPERT WILDT

The Sphere of Sacrobosco and its Commentators: by LYNN THORNDIKE. Pp. x, 496. Chicago, 1949 (The University of Chicago Press, $10.00).—The Sphere of Johannes de Sacrobosco was used as an elementary manual of astronomy in the European Universities from the thirteenth until the seventeenth century. Its presentation of the elements of Ptolemaic astronomy has a lucidity that caused it to hold its place until the Copernican system was established.

A detailed introduction which contains abundant historical notes and details on manuscripts and early printed editions of the Sphere precedes the critical edition of the Latin text and its English translation. The remainder of the book contains various commentaries in Latin together with an English translation of the commentary of Robert Anglicus.

While the work is primarily of value to specialists in the field of the history of science and the history of scholasticism, it may be recommended to every teacher of elementary astronomy. Both the Sphere and the additional material presented serve as an excellent illustration of the astronomical thinking during the Middle Ages.

DIRK BROUWER

Abriss der Geologie (Emanuel Kayser) by PROF. DR. ROLAND BRINKMANN. Vol. 1, Allgemeine Geologie, 282 pages; vol. 2, Historische Geologie, 355 pages. Stuttgart, 1940 and 1948 (Ferdinand Enke. Vol. 1, 25 D M., vol. 2, 17 D.M.).—This new edition—the sixth—of Kayser's "Abriss" has little in common with the older ones but the title and the general approach to the treatment of the subject matter. It is much less a text than a reference book. The student is not lectured to but presented with source material, charts, tables, and well chosen illustrations. As a source of information, the book is extremely valuable, especially on European data, as well as in philosophical outlook. Rich in literature references, it conveys an up-to-date picture of the status of geological work. Not competing with standard American texts, it will prove an extremely valuable and easily accessible source of information for advanced students and teachers of geology.

Volume 1—"general geology,"—is subdivided into exogenetic and endogenetic dynamics.

Solar energy keeps the exogenetic cycle in motion, thus air and water circulation, and climatic zones are discussed in the first part.

Physical and chemical weathering follow. Ground water and running water are accompanied by only a brief discussion of valley development, river systems, and sedimentation. Physiographic evolution, land forms, glaciation, weathering, soil formation, deserts, and wind action are discussed from the point of view of dominating climatic conditions.

Part 3 deals with shape and inorganic and organic content of oceans; and part 4 is devoted to oceanic regions, coastal zones, shelfs, and deep sea. The fifth part is a discussion of diagenesis and sedimentary rocks. The latter are listed in a systematic table together with equivalent mineral resources. This approach to exogenetic dynamics differs fundamentally from the viewpoint of standard American texts, and provides a continuous thread to the story.

Endogenetic dynamics comprises 4 parts: Tectonics, Magma, Regional Metamorphism, and Structure and movement picture of the globe in 12 chapters.

Tectonics deals with crustal movements, architecture, and structure Earthquakes (seismometry and the earth's interior) are recent tectonics and may result in abnormal positions of bedded rocks.

Structural geology includes folding, jointing, morphology and structure, crustal movements, and regional distribution of structures. An example of statistical analyses using the equal area net shows the approach to detailed tectonic analysis but the general emphasis is on the regional distribution of tectonic types.

Architectural forms differ where magma is involved either near the surface (volcanism) or at depth (plutonism). The magmatic rock is the product of either. Mineral resources related to igneous phenomena conclude this part. Regional metamorphism is treated briefly but adequately with reference to the most important contributors and their work.

A general review of earth physics and geophysical methods concludes the first volume and deals with density, isostasy, geochemistry, geotectonics and geotectonic theories.

Volume 2—"historical geology"—opens with methods and results of geologic time-measurement. "Erdgeschichte" proper begins in the "Grundgebirge" or basement complex, whose distribution and subdivision are briefly outlined.

From the Cambrian to the Quaternary, 10 chapters describe the geologic column systematically. Each chapter contains introductory material, a discussion of the distribution of formations over the whole world, their paleontological content, climatic conditions, facies, crustal movements, short summary, and literature references. A large correlation chart accompanies each chapter and permits comparison on a world-wide basis. Palaeogeographic maps illustrate local distribution as well as world-wide correlations. A large amount of information is concentrated into each chapter.

The systematic part is followed by two summarizing chapters on evolution—with an easily comprehended chart (p. 310)—and on the earth's history.

This volume is also more a source than a text. It contains no especially attractive titles and no catching story, but includes the basic facts in a systematic and easily available form. It may well prove to be the most useful key to European stratigraphy.

The second volume appeared in 1948 and thus the paper is not equal to that in the first volume (1940), but all illustrations are very good, and the printing is up to the best tradition of the publisher.

Author and publisher can be congratulated upon completion of this work under the most adverse conditions imaginable.

ERNST CLOOS

Introduction to Atomic Physics; by OTTO OLDENBERG. Pp. xiii, 373. New York, 1949 (McGraw Hill, $5.00).—This book describes the elements of atomic physics to students who have taken a one year introductory physics course. The arrangement of subject matter is somewhat unusual and gives evidence of the author's concern for the difficulties of teaching the subject to beginners. The treatment conveys not merely a knowledge of the facts of atomic physics, but an understanding for the intimate connection between experiment and theory. From the point of view of integration the book is one of the best I have seen. It is recommended strongly to anyone who wishes a balanced view of modern physics. The author is to be congratulated on his skillful use of historical and methodological material. H. MARGENAU

Foundations of Modern Physics; by THOMAS B. BROWN. Pp. xvi, 391. New York, 1949. Second Edition of a book first published in 1940 (John Wiley and Sons, $5.00).—Like Oldenberg's, this book covers the realm of modern atomic physics. It is essentially a serial description of experimental facts, simply presented and easily placed within the grasp of the beginner. It is less successful, in this reviewer's opinion, in integrating the conceptual and experimental components of modern physics than the book described in the preceding review. H. MARGENAU

The Science and Engineering of Nuclear Power. Vol. II; CLARK GOODMAN, Editor. Pp. 317. Cambridge, Mass., 1949 (Addison-Wesley Press, $7.50).—The first volume with this title was reviewed in the March 1948 issue of this JOURNAL None of the remarks made by the present writer in that review about the transcending importance of our newly gained knowledge of how to release in a controlled manner and on a large scale the energy stored in the nuclei of atoms need be repeated. Like the earlier book, this volume is based upon a series of seminars held at the Massachusetts Institute of Technology. It is interesting to observe how much can be told about nuclear reactors despite security regulations.

Whereas Volume I was primarily a general survey of the fundamental scientific facts about nuclear chain reactors, the contributors to Volume II write on specific aspects in some detail. Pile theory is discussed by F. L. Friedman, A. M. Weinberg, G. Placzek, H. Soodak and F. de Hoffman. With respect to nuclear engineering, there are chapters on heat removal from nuclear reactors, rockets and other thermal jets using nuclear energy, the shielding of nuclear reactors, effects of radiation on materials, and pile materials (metals, alloys and compounds). There is a valuable chapter on the production of "radionuclides" (radioactive atoms) by J. W. Irvine, Jr., and another on "health physics—instrumentation and hazard evaluation" by R. D. Evans. The book concludes with an interesting chapter by the editor on future developments in nuclear energy.

An appendix giving surveys of the unclassified literature on the energy release of the fission process, the electromagnetic methods of separating isotopes, and the number of secondary neutrons per fission should be very useful. A Segré chart summarizing the principal properties of the known atomic nuclei is also included.

It is obvious that just about all that can be written today about the physics of nuclear chain reactions and the engineering problems connected with the peace-time development of nuclear power is included in these two volumes. The discussions are authoritative and designed for the well-trained scientist. These books are a great contribution to the needed dissemination of the facts about nuclear energy.

W. W. WATSON

Botanik der Gegenwart und Vorzeit in culturhistorischer Entwickelung. Ein Beitrag zur Geschichte der abendländischen Völker. Von Karl F. W. Jessen, 1864 (Pallas, Vol. 1) Pp. xxii, 495. (Waltham, Mass.: Chronica Botanica Co.; New York City: Stechert-Hafner, Inc., 1948, $6.00).—Histories inevitably do more than provide chronicles of the events of which they treat, for they reflect the main stream of ideas and beliefs of the periods in which they were written. The spectacular success of the great "History of Botany" by Julius Sachs, published in 1875, was not entirely due to the fact that it came from the pen of one of the outstanding botanists then active. Its lucid, dynamically critical, and eminently personal style, applied to botanical development during its later, richest phases, fell into line remarkably well with that "heroic" spirit in nineteenth century science which for many signified the commencement of a new era, an age enlightened and dominated by science and reason. The work of Sachs naturally overshadowed the somewhat earlier treatises of his contemporaries, such as Ernst Meyer and Karl F. W. Jessen.

Our thanks are due to Frans Verdoorn (Chronica Botanica Company) for the excellent offset reprint edition (in German) of Jessen's rare original book which was published in 1864. This volume, printed on fine paper, is the first in a series of re-issues of classical biological works. The book should be of interest and value to both professional historians of the natural sciences and medicine and to other biologists with historical and humanistic leanings. In many respects Jessen's book supplements that of Sachs.

The history by Sachs begins with the early sixteenth century while a large portion of Jessen's book (the first ten chapters) is devoted to the beginnings of botanical science among the ancients and during the Middle Ages. Against the background of general cultural, religious, and scientific development, Jessen traces the gradual changes in botanical knowledge among the early Indians, Egyptians, and Babylonians, the Greek and the Romans, and during the early Christian era in Eastern and Western Europe. The contributions of Arabic science to botany are followed by a discussion of the beginnings of scientific activity in relation to cultural centers and schools in Western Europe. This leads approximately up to the period where Sachs begins. In the ten chapters which follow are notable sections dealing with the development of systematic botany from the early beginning to Endlicher as well as with that of agriculture and horticulture; the concluding sections discuss the general progress in natural and botanical sciences during the eighteenth and nineteenth centuries. Three appended tables illustrate the development of systematic botany, the number of botanical works printed, and notable botanical exploratory travels.

The book supplies much factual information in a field not easily accessible to most; many data and names are included, yet this does not interfere with pleasant reading. Thus in several respects the book is more than a history of botany. Its historical, philosophical, and cultural digressions demonstrate to the present reader the state of general humanistic knowledge which a scientist was able and inclined to acquire and put to good use one hundred years ago. In contemporary college teaching of botany, with its increasing accent on the role of plants in civilizations, Jessen's work should be of considerable value.

ROBERT BLOCH

The Avian Egg; *by* ALEXIS L. ROMANOFF and ANASTASIA J. ROMANOFF. Pp. 918; 424 figs. New York, 1949 (John Wiley & Sons, $14.00).—The authors of this work, recognizing that their title has almost universal symbolic connotations, use a brief preface to define their subject as 'an attempt to compile all the facts known about the bird's egg' . . . Further· 'We have limited our discussion to the egg before activation of life within it.' Thus in the controversy as to whether the hen precedes the egg or vice versa, their position is squarely for the former view: the egg as end-product. As is stated farther on (p. 175), 'Through the centuries, classical works have described in detail the development of the embryo within the hen's egg, but little attention has been given to the process by which the egg itself is formed.'

In developing this view, approximately one-third of the ample volume is devoted to the subject of egg formation; another third to the physico-chemical and biological (i.e. biochemical) properties of the egg itself; and the last section to its nutritional and industrial uses. In spite of the efforts of the authors to justify their choice of adjectives in the title, it emerges clearly that our precise knowledge of structure and chemistry depends on the egg of exactly one avian species: viz. that biological monstrosity, the domestic fowl. A tremendous amount of factual material has been classified and re-presented with admirable simplicity and clarity of expression: no little merit in a work of this sort. The efforts of the senior author in re-drawing graphs and constructing tables and diagrams contribute considerably to the unity of tone. If some of the section frontispiece diagrams appear a little naive to a professional scientist, they may nevertheless be extremely helpful to a less sophisticated reader.

The bibliography is most impressive, covering European (including of course Slavic), American, and even Asiatic literature in a wide variety of fields. One discovers citations of periodicals ranging from *Quick Frozen Foods* and the *Annales des falsifications et des fraudes*, to the *Proceedings* of the Royal Society and the *Atti* of the *Accademia dei Lincei.* Thus a comprehensive review of technical and scientific literature is achieved: a most useful compilation.

Professor and Mrs Romanoff are to be congratulated on producing a book which is not only comprehensive within its stated field, but readable. This reviewer would, however, like to offer two suggestions for a future edition: she was most disappointed to find no discussion of the problem of the cuckoo egg; and she would like to see the final anthropological section considerably expanded. She also suggests that the last plate be done over in colour as a frontispiece. Eggs are not food for the body alone.

D. RUDNICK

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Physics in the Modern World; by Henry Semat. New York, 1949 (Rinehart & Co. Inc., $5 00)

U. S. Geological Survey Bulletins as follows: 940 Strategic Minerals Investigations, 1943. Price $ 05. 944-C Geology and Ore Deposits of Boise Basin, Idaho, by A L Anderson 945. Strategic Minerals Investigations, 1944 Price $ 05 954-E Manganese Deposits of the Talamantes District near Parral, Chihuahua, Mexico; by I. F Wilson and V. S Rocha Price $ 50. 959-D. Geophysical Abstracts 135 October-December, 1948 (Nos. 10473-10736); by V. L. Skitsky and S T. Vesselowsky. Price $.25

960-A. Barite Deposits of a Camamú Bay, State of Bahia, Brazil; by A J. Bodenlos. Price $ 75. 960-B Antimony Deposits of Soyatal District State of Queretaro, Mexico; by D E White Price $1.00 960-C. Geology of Tungsten Deposits in North-Central Chile; by J F. McAllister and Carlos Ruiz F. Price $1.75. 960-E Geology of Huahuaxtla Mercury District, State of Guerrero, Mexico; by David Gallagher and Rafael Perez Siliceo. Price $ 55. 960-D Tin-Bearing Placers near Guadalcazar, State of San Luis Potosi. Mexico; by Carl Fries, Jr, and Eduardo Schmitter. Price $ 50. Professional Papers as follows: 199-B Mollusca from the Miocene and Lower Pliocene of Virginia and North Carolina. Part 2. Scaphopoda and Gastropoda; by Julia Gardner. Price $ 70 215. Geology of the Southern Guadalupe Mountains, Texas; by P. B. King Price $3 25.

214-C Paleocene Fresh-Water Mollusks from Southern Montana; by Teng-Chien Yen Price $.15.

Bulletin 954-F Manganese Deposits of Mexico; by P. D. Trask and J. R. Cabo, Jr. Price $ 35

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PREFERRED ORIENTATION OF CALCITE IN YULE MARBLE

FRANCIS J. TURNER

ABSTRACT. A detailed account of the fabric of Yule marble is given for later comparison with the fabric of the same rock artificially deformed Preferred orientation of the following crystal directions is described: optic axis; visible $\{01\bar{1}2\}$ lamellae; edges of intersecting $\{01\bar{1}2\}$ lamellae; edges between $\{01\bar{1}2\}$ and $\{10\bar{1}1\}$, edges between $\{01\bar{1}2\}$ and $\{0001\}$, optic axis within definite twin lamellae. The nature and possible origin of space-lattice orientation of the calcite grains is discussed. It is shown that many lamellae of the form $\{01\bar{1}2\}$ are probably due to translation rather than to twin gliding. Many lamellae are believed to have formed as a result of late relatively mild deformation subsequent to that responsible for the observed space-lattice orientation expressed by optic axes diagrams.

INTRODUCTION

IN a series of compression tests carried out at high confining pressures, Griggs (1936; 1940) succeeded in shortening marble and limestone columns by as much as 30-50%. He also recorded briefly some of the changes observed in the microscopic fabric of one of the test materials, viz. a pure marble from a quarry at Yule Creek, Colorado, owned by the Vermont Marble Company (see also Knopf, 1943). The committee on experimental deformation of rocks, set up under chairmanship of Mrs. E. B. Knopf by the National Research Council, has urged that this program of research be continued (Knopf, 1946); and as a preliminary step Knopf (1949) has recorded the results of a petrofabric analysis of Yule marble. Since this latter work raised or left unanswered several questions as to the state of preferred orientation of calcite in the marble, and the crystallographic identity of deformation lamellae in individual grains, Mrs. Knopf suggested to the present writer that he investigate the fabric of the same rock independently and in as full detail as possible.

The original block of marble which furnished the test cylinders for the experiments of Griggs, 1936-1940, was a geographically oriented block collected by E. B. Knopf from the

Yule quarry. The sections for the present study were cut from a second block of Yule marble, originally unoriented, which was kindly provided by Mr. Harold Ladd Smith of the Vermont Marble Company. Preliminary petrofabric analysis of this second block by Dr. H. Mikami at Yale University showed that its fabric conforms so closely with that of Griggs' original block as to allow its geographic orientation to be reliably established. Three mutually perpendicular sections were cut from the second block for the present investigation. They are approximately parallel to the three sections used by Knopf (1948) in her fabric study of the material used in Griggs' experiments, and have been labelled correspondingly.

All grains measured in the present study have been identified and recorded on three large photomicrographs showing the major portions of the three sections enlarged to 27 diameters. These photographs were made by Dr. H. Mikami, research geologist for the E. J. Lavino Company of Pennsylvania.

The writer's thanks are gratefully acknowledged to Mrs. E. B. Knopf for suggesting the problem and supplying the material, for frequent advice and consultation as the work progressed and for permitting access to unpublished results of her own work on Yule Marble. His thanks are also due to Dr. H. Mikami, Mr. Harold Ladd Smith and the Vermont Marble Company for assistance acknowledged above, and to the E. J. Lavino Company by whose courtesy facilities were granted to Dr. Mikami for preparing the photomicrographs.

For reference purposes the following details of the research material (given by Mrs. Knopf) are recorded.

Locality:	Quarry on west side of Yule Creek, 3 miles south of Marble, Gunnison County, Colorado.
Field data:	(a) Joints: (1) Main set strikes N65°E and dips 80°N.W. (2) Discontinuous joints in zones *en echelon* strike N55°W and dips 70°S.W. (b) Foliation strikes N5°E and dips approximately vertically.
Physical properties:	Pure calcite marble, free from dolomite; G = 2.7; porosity low, 0.15%; permeability fairly high (water penetrates 1.5 inches in 70 minutes).

The previous work of Knopf (1948) has demonstrated that the fabric of Yule marble is substantially homogeneous, and has given a clear picture of the state of preferred orientation of the optic axis and of all visible lamellae (whether due to twinning, translation or cleavage) in the component grains of calcite. The purposes of the present paper are:

1. to check the data recorded by Knopf;
2. to identify the crystallographic orientation of visible lamellae in calcite grains;
3. to record in as full detail as possible the state of preferred orientation of calcite in Yule marble, for subsequent comparison with fabrics of artificially deformed specimens of the same rock;
4. to record such additional data as might throw light on the mode and time of origin of visible lamellae in calcite of Yule marble.

The first part of the paper is descriptive, and includes an outline of the methods used, followed by a record of measured fabric data. Then follows a section in which the writer's tentative interpretation of the data is put forward.

LABORATORY PROCEDURE

Record of Measurements. As seen by microscopic comparison of the three sections, the single plane of foliation is defined by dimensional subparallel alinement of interlocking irregularly discoidal grains of calcite whose smallest dimensions average between one half and one third the greatest dimension. The three mutually perpendicular oriented sections were cut approximately parallel (Q) and perpendicular (P and R) to the foliation, as shown diagrammatically in Figure 1. The P section is vertical, and the R section horizontal. The tendency for calcite grains to be somewhat elongated in the plane of foliation is seen more distinctly in the R section than in P, while section Q shows a slight tendency for grains to be elongated in the horizontal (NS) direction. This confirms the previous record of a faintly developed subhorizontal (NS) lineation in the plane of foliation. The axes of the calcite fabric are thus provisionally identified as follows: *a* is vertical; *b* is NS; *c* is EW (cf. fig. 1).

On each section measurements were made on all grains, regardless of shape or size, lying on a number of separate approximately linear traverses. Each grain was numbered and identified on the corresponding photomicrograph, and a full record of optical and crystallographic measurements for each numbered grain has been kept. Elemental orientation diagrams each showing the orientation of a given optical or crystallographic direction for between ten and thirty numbered grains have also been retained.

Details of traverses and of grains measured in each section are as follows:

Section P:	100 grains within an area 13mm. x 9mm.: two vertical traverses (parallel to trace of foliation), 6mm. and 8mm. long; one EW traverse (transverse to foliation) 4mm. long; one traverse including all grains within an area 4 mm. x 1 mm.
Section Q.	90 grains within an area 10mm. x 5 mm: two vertical traverses 7mm. and 10mm. long; one traverse including all grains within an area 2mm. x 2mm.
Section R:	150 grains within an area 13mm. x 10mm.: two EW traverses (transverse to trace of foliation), 12mm. and 7mm. long; one traverse including all grains in an area 5mm. x 3mm.

Since the fabric has already been shown to be homogeneous, it was thought advisable to concentrate on detailed investigation of relatively few (340) grains. In each crystal the position of the optic axis was measured and checked; all visible deformation lamellae and the more obvious cleavage directions were measured and crystallographically identified; additional but not obvious lamellae were searched for wherever within the accessible field of view; individual peculiarities such as relative grain size, bending of certain sets of lamellae, development of obvious twinning in other sets, relation of lamellae in adjoining grains and relation of lamellae to grain outlines were recorded wherever observed. Yet other crystallographic directions were determined graphically from the positions of the optic axis and the poles of lamellae or cleavages as plotted on a Schmidt projection net.

Finally elemental and collective orientation diagrams were prepared for each of the following directions, as determined in each of the three oriented sections.

1. Optic axis = vertical crystal axis.
2. Poles of visible lamellae of the form $\{01\bar{1}2\} = e$.
3. Edges $[e:c]$, i.e. edges between visible lamellae $\{01\bar{1}2\}$ and the basal pinacoid $\{0001\}$.
4. Edges $[r:r'] = [r:e]$ i.e. edges between visible lamellae $\{01\bar{1}2\}$ and either of the adjoining unit rhombohedron faces $(10\bar{1}1) = r$ or $(\bar{1}101) = r'$.
5. Edges $[e:e']$, i.e. edges between visible faces of the form $\{01\bar{1}2\}$
6. Edges $\frac{\perp [e:e']}{(0001)}$, i.e. the line drawn in the basal pinacoid perpendicular to the edge between visible faces of the form $\{01\bar{1}2\}$.
7. The optic axis within each observed $\{01\bar{1}2\}$ lamella, assuming that the latter is due to twinning.

Additional diagrams were also prepared for optic axis, pole of $\{01\bar{1}2\}$ lamella, new position of optic axis, and glide-line $[r:r'] = [e:r]$ in those grains in which twinning on $\{01\bar{1}2\}$ could be demonstrated beyond doubt.

Orientation diagrams depicting data for any one type of crystallographic direction (e.g. normals to lamellae) were prepared separately for each of the three sections examined. For detailed comparison, and for preparing collective diagrams based on measurements in more than one section, it was found more satisfactory to rotate the plotted points individually through 90°, than to rotate contoured diagrams. This procedure has therefore been followed throughout, wherever rotation seemed desirable.

Measurement of the c *crystal axis* $[0001]$. Several difficulties attend location of the c crystal axis of a calcite grain with a universal stage. The extreme birefringence of calcite makes it difficult to differentiate quickly between fast and slow directions in a given grain unless the section is thin. Moreover, for some orientations, especially where the c axis is inclined at a small angle to the plane of sections, the grain may remain in extinction through as much as 10° of tilt. An added difficulty in the case of Yule marble arises from the distinctly biaxial condition of many grains ($2V = 0° — 15°$). The procedure outlined below was found generally satisfactory in minimizing errors arising from these causes. Repeated checking of angles between the poles of measured lamellae (or cleavage cracks) and the *c* axis, show that the position of the latter can always be located within two or three degrees, i.e. sufficiently accurately for requirements of fabric analysis.

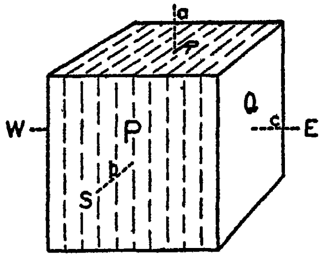
(1) The section is mounted between glass hemispheres having refractive index 1.557, using colorless mineral oil ("albolene") as mounting oil. All measurements of optic axes are made with the optic axis in the E.W. vertical plane, so that light (vibrating N.S.) is transmitted entirely as the ordinary ray ($\gamma = 1.659$). The stage may be tilted up to 60° without total reflection; but all tilts must be corrected for difference in refractive index of hemispheres and calcite. Thus a reading of 40° is reduced to 37°; 25° is reduced to 23°, and so on.

(2) With both axes of tilt at the zero position, the calcite grain is first rotated on the innermost stage axis to extinction, with the fast direction (trace of the optic axis) E.W. [In identifying the fast direction with the aid of a quartz wedge, the position of compensation may usually be recognized clearly if the iris diaphragm in the objective is first stopped down. The test is not decisive if the section is thick and is cut nearly parallel to the optic axis. In this case, however, the apparent relief of the section is obviously higher when the slow direction is N.S. than when it lies E.W.]

(3) The stage is now tilted through about 30° on E.W. and the section once more becomes illuminated (unless the optic axis already coincides with the E.W. stage axis). Tilting on N.S. to extinction now brings the optic axis either parallel to E.W. or vertical. The latter case is readily distinguished by the manner in which the order of interference color falls through first order greys to the extinction position. If necessary, it may be confirmed by tilting back on E.W. to the zero position, unclamping the stage, and rotating on the main vertical axis (microscope axis). The grain, of course, remains dark only if the optic axis is vertical.

(4) A useful check may be obtained by measuring three or more directions of ordinary ray vibrations. The stage is

set with E.W. at zero and is tilted about N.S. to 20° from the left. The section is now rotated on the innermost vertical axis until it reaches extinction with the slow direction (ordinary ray) E.W. This procedure is repeated independently with N.S. first at zero, and then tilted 20° from the right.



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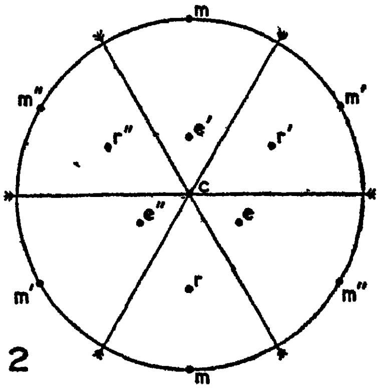

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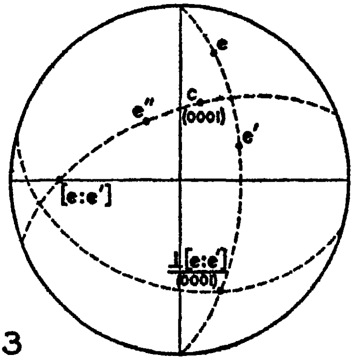

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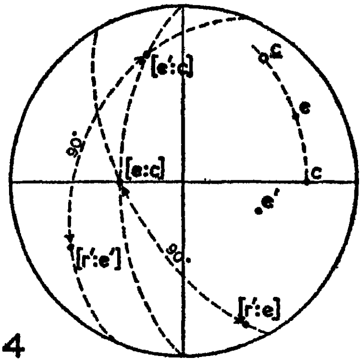

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Figure 1. Section block of Yule marble showing orientation of the three sections P, Q, and R with respect to geographic coordinates and fabric axes *a*, *b*, *c*.

Figure 2 Equal-area projection showing relative positions of poles of faces of the forms $\{10\bar{1}0\}$ = m, $\{10\bar{1}1\}$ = *r*, $\{01\bar{1}2\}$ = *e*, and $\{0001\}$ = *c*, in calcite.

Figure 3. Equal-area projection showing relative positions of poles of faces of forms $\{01\bar{1}2\}$ = *e*, and $\{0001\}$ = *c*, the edge [e : e′] between two faces of the form $\{01\bar{1}2\}$; and the normal to this edge lying in the face (0001), i.e. $\frac{\perp [e : e']}{(0001)}$ Drawn for calcite.

Figure 4. Equal area projection showing edges [*e* : *c*] and [*e′* : *c*] between (0001) and faces of form $\{01\bar{1}2\}$; edges [*r′* : *e*] and [*r′* : *e′*] between faces of the forms $\{01\bar{1}2\}$ and $\{10\bar{1}1\}$; optic axis $\underline{c}$ of lamella twinned on *e*. Drawn for calcite.

The three corresponding points plotted on the projection net should lie on a great circle whose pole is the point of emergence of the c axis. The writer has found this procedure somewhat the more satisfactory of the two in cases where the optic axis is inclined at a low angle to the plane of the section, or where the grain is either biaxial or strained (as indicated by undulose extinction.)

Measurement of lamellae and cleavage. A lamella or cleavage plane is observed in the normal manner by bringing its trace parallel to either cross-hair and tilting to the vertical position about the corresponding axis of the stage. The other titling axis is kept at the zero position. Lamellae of the form $\{01\bar{1}2\}$ are more readily observed between crossed nicols, but cleavages $\{10\bar{1}1\}$ are more sharply defined when seen in ordinary light. Partial stopping down of the iris diaphragm in the objective may sharpen definition.

Wherever an individual lamella shows alternate illumination and extinction on rotation of the whole stage, and the lamella and the main portion of the grain give symmetrical extinction angles, the lamella is recorded as clearly due to twinning. In Yule marble the majority of lamellae of the form $\{01\bar{1}2\}$ are not obviously twinned. When tilted into the vertical plane they appear as hair-sharp lines, rather than as lamellae. But they may usually be distinguished from cleavages $\{10\bar{1}1\}$ by their greater individual continuity and by a tendency to remain clearly visible even when tilted at angles of as much as 30° from the vertical plane.

Each set of cleavages or lamellae is identified with certainty as belonging to either $\{10\bar{1}1\}$ or $\{01\bar{1}2\}$ by plotting its pole and measuring the angle between the pole and the c axis. The following interfacial angles (measured between poles of faces) are of use in this connection (cf. fig. 2).

$c \wedge r$, $(0001) \wedge (10\bar{1}1) = 44\frac{1}{2}°$
$c \wedge e$, $(0001) \wedge (01\bar{1}2) = 26\frac{1}{4}°$
$e \wedge e'$, $(01\bar{1}2) \wedge (1\bar{1}02) = 45°$
$r \wedge r'$, $(10\bar{1}1) \wedge (\bar{1}101) = 75°$
$e \wedge r$, $(01\bar{1}2) \wedge (10\bar{1}1) = 38°$

Note that the interfacial angles between a face $(01\bar{1}2)$ and the adjacent faces, $(10\bar{1}1)$ and $(\bar{1}101)$ are 38°; but the third unit rhombohedron face $(0\bar{1}11)$ is co-zonal with $(01\bar{1}2)$ and the intervening face (0001); so that the angle $(01\bar{1}2) \wedge (0\bar{1}11)$ is $70\frac{3}{4}°$.

When one or two faces of the form $\{01\bar{1}2\}$ have been measured and plotted, the positions of the remaining faces or face of the same form may be located on the projection. The innermost rotation axis and one tilting axis of the stage are now set at appropriate angles and the section is searched, to make sure that lamellae parallel to the unrecorded face or faces have not been overlooked.

Location of edges $[e:e']$ *and* $\frac{\perp[e:e']}{(0001)}$. Where two sets of lamellae e and e' belonging to the form $\{01\bar{1}2\}$ have been measured in one grain, their line of intersection is located graphically on the projection net. Its projection ($[e:e']$ in fig. 3) is the pole of the great circle which includes the poles of both lamellae e and e'. The direction within (0001) which is normal to the intersection $[e:e']$, here designated $\frac{\perp[e:e']}{(0001)}$, is the pole of the great circle which includes $[e:e']$ and the pole of (0001). It is also parallel to one of the three horizontal crystal axes, i.e. to the edge between the basal pinacoid and the third rhombohedron e'' of the form $\{01\bar{1}2\}$. These relations are shown in figure 3.

Location of edges $[e \cdot c]$ *and* $[e:r]$. For each measured set of faces of the form $\{01\bar{1}2\}$, the edge of intersection with the basal pinacoid (0001) is located graphically as the pole of the great circle on which the c axis and the pole of the lamella both lie (fig. 4). The glide direction for normal lamellar twinning, namely $[r:r'] = [e:r]$, is normal to the edge $[e:c]$ and lies within the plane e. It may be located by measuring off an angular distance of 90° from the point $[e:c]$ along the great circle representing the e lamella. Where more than one set of lamellae $\{01\bar{1}2\}$ is developed in one grain, the observed angles between corresponding edges $[e:c]$ and $[e':c]$, or $[e:r]$ and $[e':r]$ afford a useful test as to accuracy with which lamellae and optic axis were located in the first instance. The true values are:

$$[e:c] \wedge [e':c] = 60^\circ$$
$$[e:r] \wedge [e':r] = 78^\circ$$

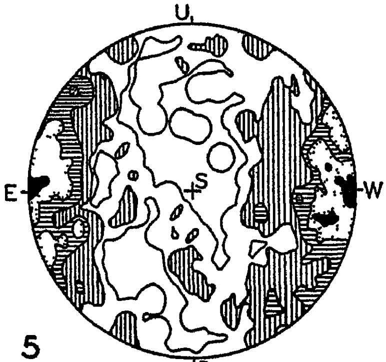
Location of c *crystal axis in twin lamellae.* In Yule marble twin lamellae are seldom wide enough to allow the position of the optic axis within a lamellae to be located directly. Since, however, the crystallographic directions of a lamella and those of the main crystal in which it occurs are related as by reflection across the twin plane $\{01\bar{1}2\}$, the optic axis in the lamella may be located graphically. Rotate the projection until the pole of the lamella and the projection of the *c* axis of the main crystal lie on a great circle. Then the optic axis within the twin lamella emerges at a point 26¼° from the lamella pole on the same great circle (fig. 4).

RESULTS OF PETROFABRIC ANALYSIS OF YULE MARBLE

Preferred orientation of c *axis* [*0001*]. The orientation of all vertical crystal axes measured (340) is shown in a collective diagram, figure 5. Corresponding elemental diagrams showing the same data plotted separately for each of the three sections are reproduced for comparison (figs. 6, 7, 8). The most significant feature of these diagrams is a constant but somewhat dispersed maximum coinciding with the pole of the foliation. Submaxima are on the whole unimportant, and there seems to be no indication of a girdle pattern in the fabric. A count of the plotted axes shows that 52% lies at angles ranging from 55° to 90° to the plane of foliation. Within this sector the axes are fairly evenly distributed except for a slightly denser concentration at approximately 90° to the foliation. Separate plots of axes of smallest and largest grains respectively show no significant differences.

There is general agreement between these diagrams and those previously published by E. B. Knopf (Griggs, 1940, p. 1006, fig. 1). The latter show the main axes maximum divided into submaxima which are asymmetrically situated with respect to the pole of the foliation; moreover there is a weak broken girdle in Knopf's diagrams. Neither of these features is reproduced in the writer's diagrams.

Preferred orientation of lamellae {*0112*}. Since lamellae or cleavages inclined at angles of less than about 35° to the plane of the section are either invisible or cannot be located accurately, there is a central blank area or blind spot in any orientation diagram constructed for poles of lamellae observed in only one section. In Yule marble a fairly complete picture of the state of preferred orientation of lamellae $\{01\bar{1}2\}$ may be obtained from any section (e.g. P and R) cut normal to the foliation. In these, the *c* crystal axis in most grains lies close to the plane of the section, so that all three sets of lamellae of the form {01$\bar{1}$2} fall within the field of visibility. Of the grains measured in sections P and R, 73% have their *c* axes inclined at angles of less than 30° to the plane of the section, and so satisfy this condition completely. Only 12% have their optic axes so steeply inclined (at 50°-90°) to the section, that at least one set of {01$\bar{1}$2} lamellae must lie outside the field of accurate measurement. But even in these, lamellae may be located indirectly, provided they become visible at high angles of tilt. The trace of such a lamella is brought parallel to the N.S. crosshair by rotating the section



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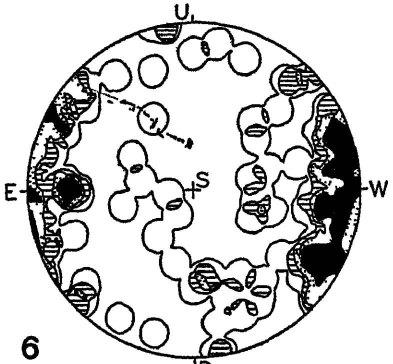

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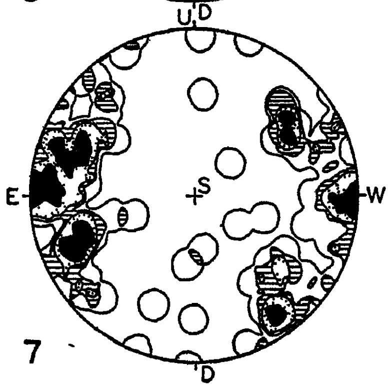

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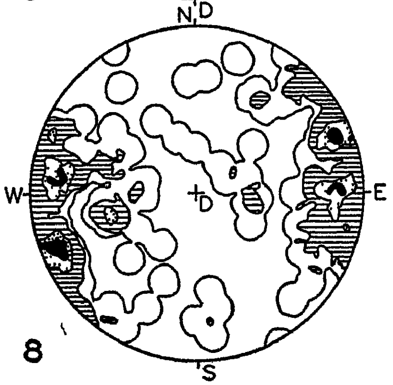

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Figure 5. [0001] = *c* axis for 340 grains measure in sections P (100), Q (90) and R (150), and plotted on P. Contours at 4%, 3%, 2%, 1%, 0.3%, per 1% area.

Figure 6. [0001] for 100 grains in section P. Contours at 4%, 3%, 2%, 1%.

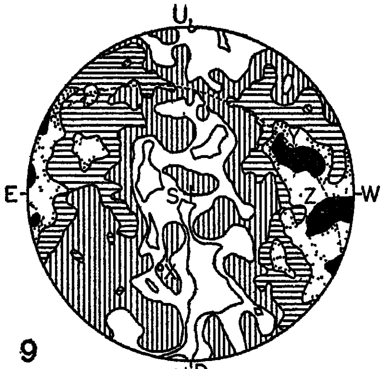
Figure 7. [0001] for 90 grains in section Q, rotated into plane of section P. Contours at 4%, 3%, 2%, 1%.

Figure 8. [0001] for 150 grains in section R. Contours at 6%, 4%, 2%, 0.7%.

on the innermost axis and tilting on the N.S. axis to a high angle, with the E.W. axis set in the zero position. The pole of the lamella may now be plotted in the appropriate position 26¼° distant from the *c* axis of the same grain.

Figure 9 is a collective diagram based on all measured lamellae (a total of 571) in the three oriented sections. Of these, 188 were measured in section P; 136 poles measured in Q and 247 in R were rotated into the plane P. The collective diagram should give an adequate picture of preferred orientation of $\{01\bar{1}2\}$ lamellae developed in this specimen of Yule marble. Its main feature, viz, a broad sector of concentration around the pole of the foliation, is reproduced in the corresponding elemental diagrams (figs. 10-13), if allowance is made for the blind spot surrounding the pole of the foliation in figure 11. A faint but distinct girdle pattern may be discerned particularly in figures 9, 12 and 13. The girdle axes (X, fig. 9) is unrelated to any of the recorded elements of the megascopic fabric. Subdivision of the sector of concentration into two main maxima (Y and Z, fig. 9) is shown independently in figures 10 and 13 and may, therefore, be significant. The lamellae diagrams of Knopf (1949) likewise show a maximum divided about the pole of the foliation. This author did not distinguish, however, between $\{01\bar{1}2\}$ lamellae and $\{10\bar{1}1\}$ cleavages. So that her lamellae diagrams are not strictly comparable with those accompanying this paper.

Of the 250 grains measured in sections P and R, 5% lack visible $\{01\bar{1}2\}$ lamellae; 30% show one set, 50% two sets, and only 15% three sets of lamellae. We have already noted that over 70% of the grains in these two sections are so oriented that all three sets of lamellae, if developed, would be accessible for measurement. It would thus appear that development of visible lamellae has been strongly selective. Consequently, figures 9 to 13 do not express completely the state of preferred orientation of the calcite *space-lattice* in Yule marble. Their pattern is determined partly by the existing preferred orientation of the lattice, but partly, too, by the selective mode of development of the measured lamellae. That this selective influence is not a random one is suggested by the distinct though weak girdle pattern and accompanying concentration of lamella poles in the upper sectors of figures 9 to 13. No such features are apparent in the corresponding axes diagrams such as figure 5.



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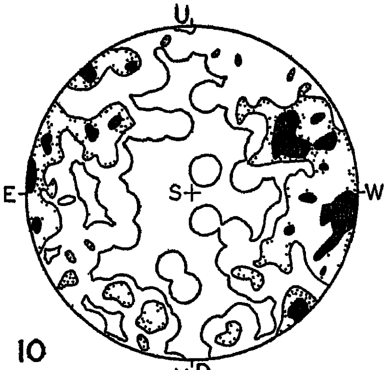

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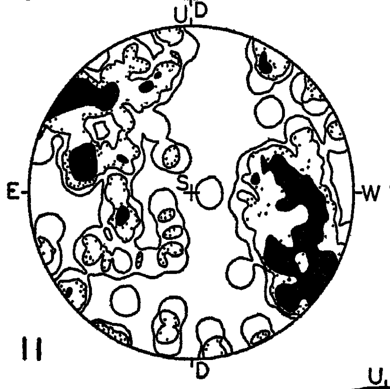

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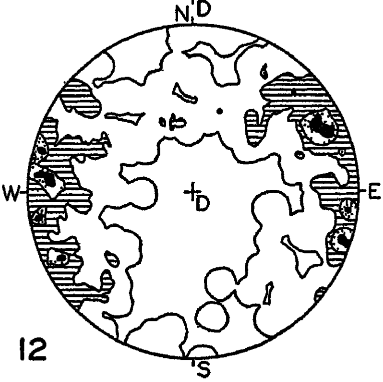

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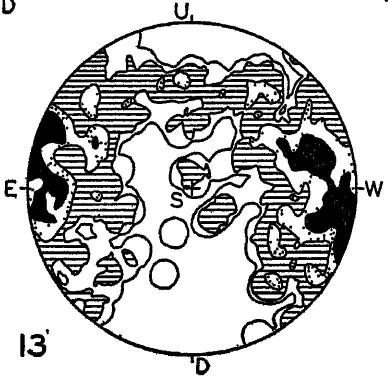

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Figure 9. Poles of $\{01\bar{1}2\} = e$: 571 lamellae measured in sections P (188), Q (136) and R (247), and plotted on P. Contours at 3%, 2%, 1%, 0.5%, 0.2%, per 1% area.

Figure 10. Poles of $\{01\bar{1}2\}$, 188 lamellae in section P. Contours at 3%, 1.6%, 0.5%

Figure 11. Poles of $\{01\bar{1}2\}$, 136 lamellae in section Q, rotated into plane of section P. Contours at 3%, 1.5%, 0.7%.

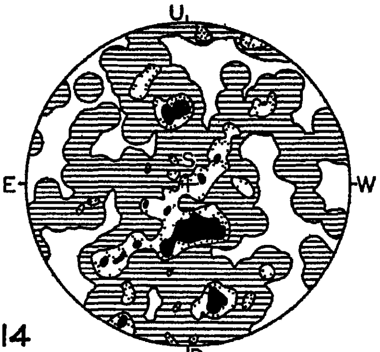
Figure 12 Poles of $\{01\bar{1}2\}$, 247 lamellae in R. Contours at 5%, 4%, 2%, 0.4%.

Figure 13. Poles of $\{01\bar{1}2\}$, 247 lamellae in R, rotated into plane of section P. Contours at 3%, 2%, 1%, 0.4%.

Preferred orientation of edges [*e* : *e*]. Any preferred orientation of the calcite lattice, in addition to the general E.-W. alinement of optic axes already noted, should be reflected in orientation diagrams for edges between $\{01\bar{1}2\}$ lamellae and basal pinacoid; for these also coincide with one or other of the three horizontal crystal axes in any given grain. The orientation of such edges for all visible lamellae in the P and R sections is shown in figures 14 and 15 respectively. The corresponding diagram for section Q is in all respects similar and has been omitted. The broad ill-defined vertical N.-S. girdle and accompanying E.-W. minimum may be correlated with the general E.-W. alinement of optic axes depicted in figure 5. But the edges are not uniformly distributed within the N.-S. girdle. In all three diagrams, more edges dip south than dip north, so that there is a greater concentration of poles in the S.-D. than in the N.-D. sector of the girdle. This condition may be correlated with the same selective development of deformation lamellae as is indicated by the lamellae diagrams.

To evaluate the possible influence of space-lattice orientation of the calcite grains upon the pattern of figure 14 and 15, the poles of all three $\{01\bar{1}2\}$ faces were located (by direct measurement or by graphic construction) in 97 out of the 100 measured grains of section P*. The state of preferred orientation of the corresponding edges, $\{0001\}:\{01\bar{1}2\}$, is shown in figure 16, in which the poles are more uniformly distributed through the N.-S. girdle than is the case in figures 14 and 15. This is strong evidence against preferred orientation of horizontal crystal axes in the calcite fabric; but it is still conceivable that a weak orientation of horizontal axes exists but is obscured by the imperfect state of alinement of vertical axes shown by the spreading maximum of figure 5. This latter possibility has been tested by plotting all three edges $\{0001\}:\{01\bar{1}2\}$ for all grains (10 in each section) whose optic axes lie within 12° of the ideal horizontal E.-W. direction which corresponds to the maximum in figure 5. As shown in figure 17, these edges are almost uniformly distributed in the N.-S. vertical plane. In two sections (P and Q) some degree of orientation may be detected; but the maximum concentrations for one section coincide with minima for the other, while in the third section (R) the distribution is random.

* In the three remaining grains, neither $\{10\bar{1}1\}$ cleavages nor $\{01\bar{1}2\}$ lamellae were visible



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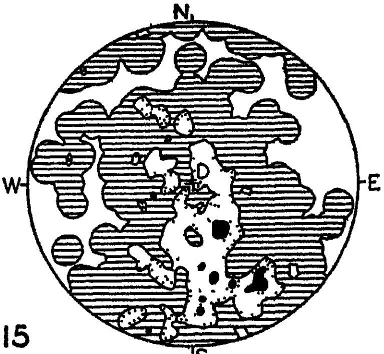

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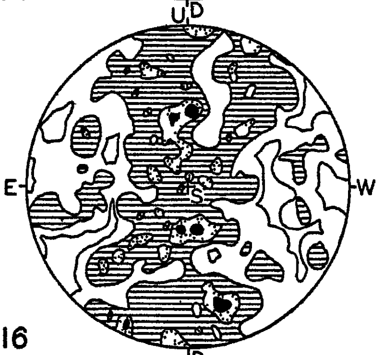

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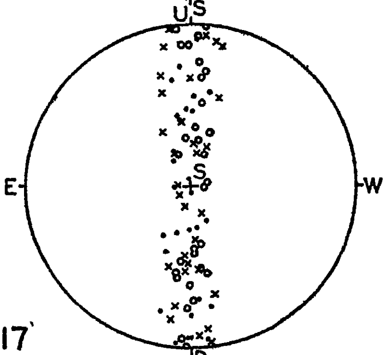

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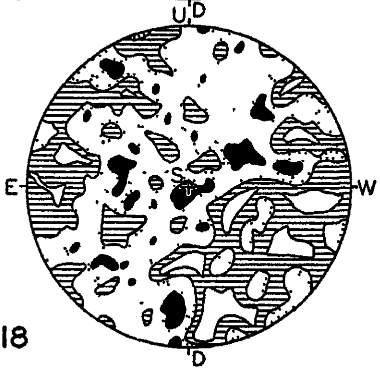

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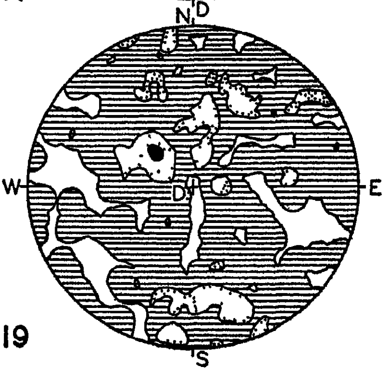

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Figure 14. Edges $[e : c]$, in 188 lamellae in P. Contours at 3%, 2% 0.5%, per 1% area.

Figure 15. Edges $[e : c]$, in 247 lamellae in R. Contours at 4%, 2%, 0.4%.

Figure 16. All three edges $[e : c]$ in 97 grains (out of a total 100) in P. Total number of edges = 291. Contours at 3%, 2%, 1%, 0.4%.

Figure 17. All three edges $[e : c]$ for 30 grains whose optic axes fall within 12° of the E.-W. horizontal direction. Dots represent poles measured in section P; open circles poles measured in Q; crosses poles measured in R; all plotted on plane of section P.

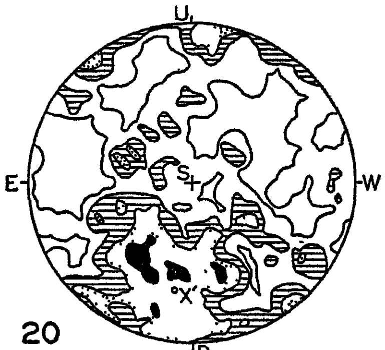
Figure 18. Edges $[e : r] = [r : r']$ in 324 lamellae measured in sections P (188) and Q (136), and plotted on P. Contours at 2%, 1% and 0.3%

Figure 19. Edges $[e : r] = [r : r']$ in 242 lamellae in R. Contours at 4%, 2%, 0.4%.

It is therefore concluded that for the rock as a whole the horizontal crystal axes of the calcite grains have random orientation, but that there may possibly be very local fields of weak orientation within the individual thin sections. The only significant preferred orientation of the calcite lattice is the E.-W. alinement of optic axes. The girdle pattern of the visible deformation lamellae and the slight concentration of their edges (intersections with $\{0001\}$) in a southward-dipping direction are directly correlated phenomena attributed to selective development of deformation lamellae. That this selection of certain $\{01\bar{1}2\}$ planes and omission of others was controlled by their respective orientations in space, is shown by the fact that the visible lamellae conform to a uniform pattern recognizable in all three sections.

Preferred orientation of edges $[e:r] = [r:r']$. The edge between the twinning plane $(01\bar{1}2)$ and either of the adjoining unit rhombohedrons $(01\bar{1}1)$ or $(\bar{1}101)$ is the glide line for twin gliding on $(01\bar{1}2)$. Orientation diagrams for this direction within visible deformation lamellae are reproduced as figures 18 (sections P and Q) and 19 (Section R). The general lack of preferred orientation is remarkable. The only consistent feature seems to be a poorly defined broad sector of minimum concentration covering the S.-W.-D. quadrant.

Preferred orientation of edges $[e:e']$. The most consistent element in the fabric so far as deformation lamellae are concerned is a marked preferred orientation of the edges common to two visible lamellae in any given grain. We have already seen that about 50% of the measured grains show two sets of lamellae, and about 15% show three sets. Of those showing three, about one-half have two well developed sets of lamellae, and one poorly defined set which was excluded for present purposes. The collective diagram, figure 20, shows the orientation of 246 edges of the type $(01\bar{1}2):(1\bar{1}02)$ measured in all three sections. Approximately 70% of these represent grains with only two sets of lamellae; 15% are edges between well developed lamellae in grains which also show a weakly defined third set (here excluded); the remaining 15% are the mutual intersections of three equally prominent sets of lamellae in a small number of grains. The consistent feature common to all three elemental diagrams (figs. 21 to 23) and to the collective diagram (fig. 20) is a strong concentration of edges in the D.-S.-E. quadrant. In other words, there is a



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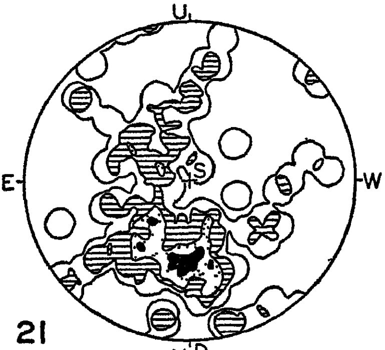

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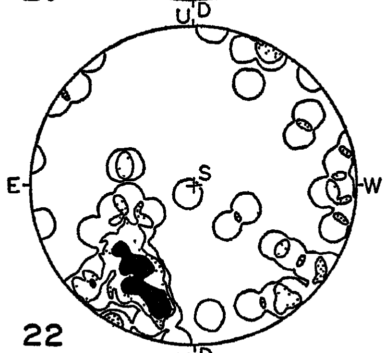

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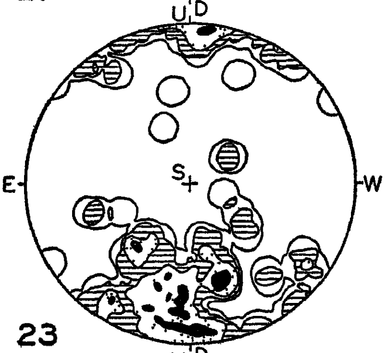

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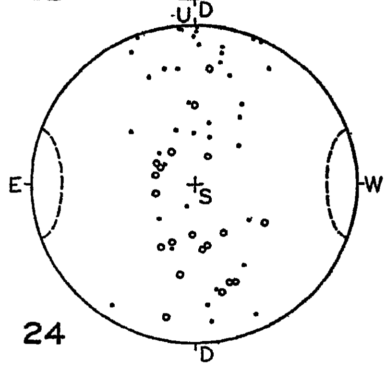

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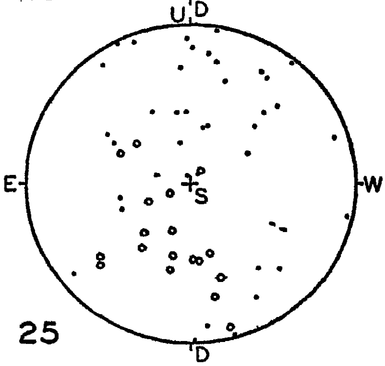

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Figure 20. Edges [*e* : *e′*] for 246 pairs of *e* lamellae measured in P (90), Q (57) and R (99) and plotted on plane of P section. Contours at 4%, 2%, 1%, 0.4%, per 1% area.

Figure 21. [Edges [*e* : *e′*] for 90 pairs of *e* lamellae in section P. Contours at 7%, 4%, 2%, 1%.

Figure 22. Edges [*e* : *e′*] for 57 pairs of e lamellae measured in section Q and rotated into plane of section P. Contours at 7%, 4%, 2%.

Figure 23. Edges [*e* : *e′*] for 99 pairs of *e* lamellae, measured in section R and rotated in to plane of section P. Contours at 6%, 4%, 2%, 1%.

Figure 24. All three edges [*e* : *e′*] for 19 grains in P, with optic axes within 20° of E.W. horizontal direction (i.e., within the broken arcs). Open circles represent edges between observed pairs of lamellae.

Figure 25. All three edges [*e* : *e′*] for 20 grains selected at random in section P. Open circles represent edges between observed pairs of lamellae.

very pronounced tendency for the edges between intersecting deformation lamellae to dip steeply in a southerly to southeasterly direction, which is also the axis of the girdle in the lamellae diagram, figure 9. This tendency is so strong that it becomes obvious in any plot of ten or fifteen edges.

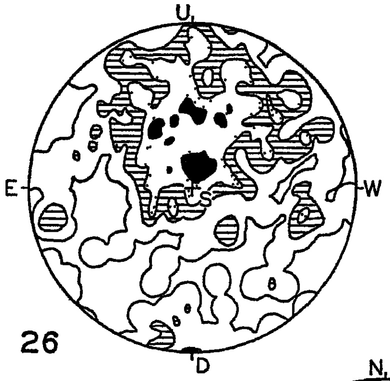
To test the possible influence of existing space-lattice orientation of grains upon the pattern of figure 20, all three edges of the type $(01\bar{1}2):(1\bar{1}02)$ were located and plotted for a selected group of 19 grains from section P, whose optic axes lie within 20° of the ideal E.-W. horizontal direction. For comparison, all three edges were also plotted for each grain in a random group (numbers 1 to 20) in the same section. Both diagrams (figs. 24, 25) show equally strong preference for edges between visible deformation lamellae to lie in the S.-D. rather than in the S.-U. sector. Pre-existing space-lattice orientation thus appears to have had no influence upon the pattern of figure 20.

Preferred orientation of $\frac{\perp [e:e']}{(0001)}$. The strong orientation of edges $[e:e']$ discussed in the previous section is accompanied by equally strong orientation of that line within $\{0001\}$ which is perpendicular to $[e:e']$ the common edge between visible deformation lamellae. This crystallographic line is here designated $\frac{\perp [e:e']}{(0001)}$. Its state of preferred orientation is illustrated by the collective diagram figure 26 which is based upon perpendiculars drawn to all $[e:e']$ edges plotted in figure 20. The corresponding elemental diagrams for grains measured in section R (figs. 27, 28) bring out the same relationship.

The line $\frac{\perp [e:e']}{(0001)}$ is also one of the three horizontal crystal axes. For example, in a grain showing two sets of deformation lamellae, $e = (01\bar{1}2)$ and $e' = (1\bar{1}02)$, the line $\frac{\perp [e:e']}{(0001)}$ coincides with the common edge between (0001) and the third rhombohedron plane $(\bar{1}012)$, i.e. with the crystal axis A_2. Figures 26 and 27 thus illustrate the state of preferred orientation of edges of the type $[e:c]$ for those $\{01\bar{1}2\}$ faces which have remained inactive during deformation of the rock. It is interesting to compare figure 27 with figure

15, which shows the orientation of edges $[e : c]$ for active $\{01\bar{1}2\}$ lamellae in the same section (R).

Preferred orientation of directions within twin lamellae $\{01\bar{1}2\}$. Measurements to be described in this section were made with a view to determining the relative parts played by twin gliding and translation gliding in the origin of visible deformation lamellae. Poles of optic axes and lamellae were plotted separately for grains in which the twinned nature of at least one lamella could be recognized with certainty. This was the case for only 50 grains (58 sets of twinned lamellae)



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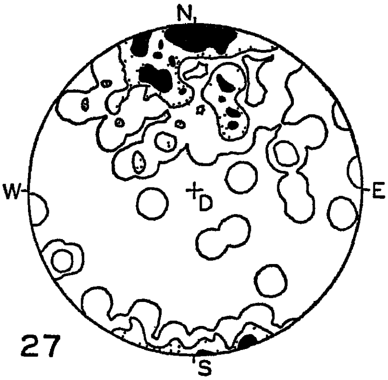

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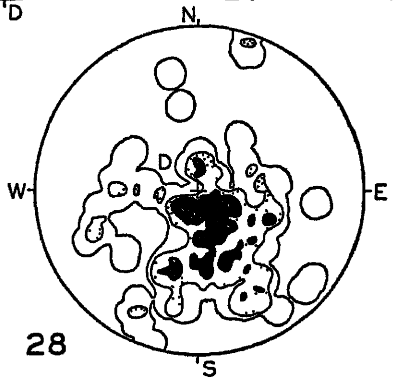

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Figure 26. Direction $\frac{\perp [e : e']}{(0001)}$ for 246 pairs of lamellae measured in P (90) Q (57) and R (99) sections and plotted on plane of P section Contours at 4%, 2%, 1%, 0 4%, per 1% area.

Figure 27. Direction $\frac{\perp [e . e']}{(0001)}$ for 99 pairs of lamellae in section R. Contours at 5%, 3%, 1%.

Figure 28. Edges $[e : e']$ for 99 pairs of lamellae in section R. Contours at 5%, 3%, 1%.

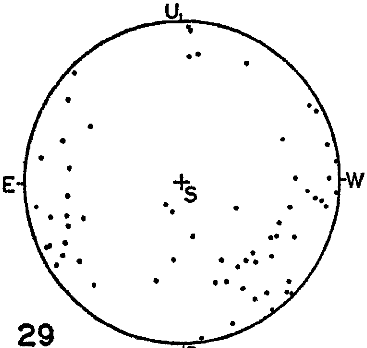
out of the total 340 grains (571 sets of lamellae) measured. Measurements were also made on 15 additional grains with unusually clear twinning (18 sets of lamellae) in Section R.

Figures 29 and 30, respectively, show the orientation of optic axes in the main untwinned portions of the 65 grains concerned, and the poles of the 76 sets of definite twin lamellae developed in these grains. Comparison with figures 5 and 9 shows that undoubted twinning is restricted to grains with a rather narrow range of orientation. Twinning seems to show a marked preference for grains whose optic axes lie in the lower half of the orientation diagram, i.e. dip gently downward, somewhat to the south of the ideal E.-W. direction which corresponds to the maximum of figure 5. The lamellae themselves are concentrated asymmetrically in the two S.-W. quadrants of figure 30, leaving the S.-E. quadrants almost unoccupied. The corresponding glide lines $[r : r']$ shown in figure 31, like those of figure 18, show no obvious regularity of orientation.

The space-lattice within the newly twinned lamellae is even more sharply oriented, as indicated by figure 32 (optic axes within 76 sets of twin lamellae). Notable features of this orientation pattern, equally apparent in all three corresponding elemental diagrams for P Q and R sections respectively, are the strong concentration of poles in the U.-S.-W. quadrant and the complementary unoccupied area in the D.-S.-W. quadrant. It is interesting to compare figure 32 with figures 33 and 34. The latter show the orientation of optic axes within all observed deformation lamellae on the assumption that these are the result of twinning. They faintly resemble figure 32 in that they show a weak maximum in the U.-S.-W. (= D.-N.-E.) quadrant. The implications of this contrast will be discussed later; but in the meantime, it may be noted as evidence in general support of the hypothesis that twinning was not involved in the formation of a high proportion of visible deformation lamellae in Yule marble.

INTERPRETATION OF PREFERRED ORIENTATION PHENOMENA IN YULE MARBLE

Nature and origin of space-lattice orientation of calcite. The space-lattice orientation of calcite in Yule marble conforms to a simple pattern. The *c* axes tend to lie at high angles to the plane of foliation which itself is defined by dimensional



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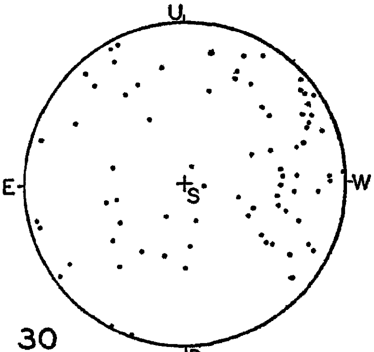

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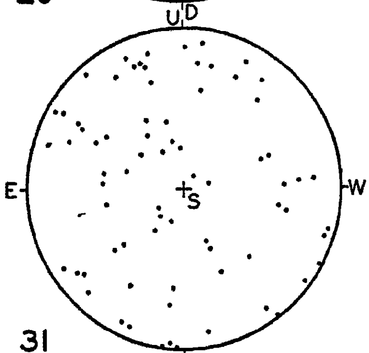

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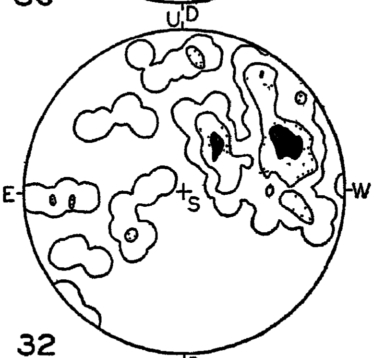

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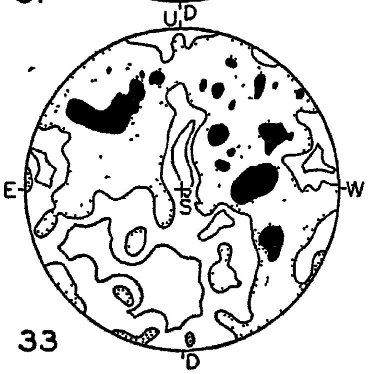

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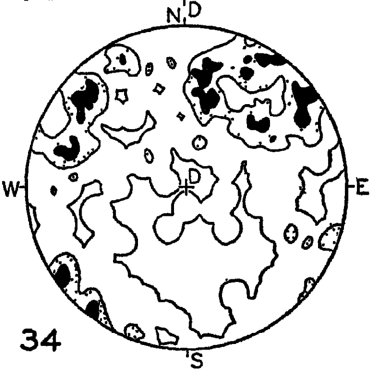

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Figure 29. Optic axes in main portions of 65 grains showing at least one set of definite twin lamellae *e*; measured in P, Q, and R sections and plotted on plane of P section.

Figure 30. Poles of 76 definite twin lamellae *e* in 65 grains of figure 29.

Figure 31. Edges [*e* : *r*] = [*r* : *r'*] for 76 definite twin lamellae *e* in 65 grains of figures 29 and 30.

Figure 32. Optic axes [0001] in 76 definite twin lamellae *e* of figure 30; measured in P (24), Q (26) and R (26), and plotted on plane of P section. Contours at 7%, 4%, 1.3%.

Figure 33. Optic axes [0001] in 324 observed lamellae *e*, assuming twinning for each set of lamellae; measured in P (188) and Q (136) and plotted on plane of P section. Contours at 2%, 1%, 0.3%.

Figure 34. Optic axes [0001] in 247 observed lamellae *e* of section R, assuming twinning for each set of lamellae. Contours at 3%, 2%, 0.4%.

parallelism of somewhat lenticular grains. Preferred orientation of no other direction in the calcite lattice has been detected.

Orientation of calcite with the *c* axes normal to a single set of visible *s*-planes has been explained in several ways:—

> (1) The experimental work of Griggs showed that simple compression of dry marble can produce a very pronounced orientation of *c* axes parallel to the direction of compression (Griggs, 1940, p. 1006, fig. 2). With this is associated equally strong preferred orientation of $\{01\bar{1}2\}$ twin lamellae, in planes symmetrically inclined at 65°-70° to the direction of compression (Griggs, 1936, p. 568).
>
> (2) Fairbairn (1941b, pp. 629-631) has drawn attention to the possibility that translation on {0001} may be an effective orienting mechanism for dolomite. Conceivably the same mechanism might be possible for calcite, in which case the *c* axes would show preferred orientation normal to to s-planes of slip, and the glide line (= one of the three horizontal crystal axes) should be oriented parallel to the slip direction within the *s*-planes.
>
> (3) In growth fabrics calcite may crystallize as tabular grains elongated at right angles to the *c* crystal axes and parallel to the direction of greatest ease of growth in the rock fabric (Sander, 1930, p. 206; Reed, 1945, p. 384). The state of orientation observed in Yule marble thus could possibly be due to mimetic preservation of pre-existing *s*-surfaces by post-tectonic recrystallization. Felkel (1929, pp. 62, 63) has described cases where recrystallized grains in deformed marble have their optic axes alined parallel to *s*-planes developed during preceding deformation.
>
> (4) Finally there is the possibility that orienting of calcite is the result of recrystallization (indirect componental movement) of the rock under simple compression. Such a mechanism would explain the fabric developed in the preliminary experiment of Griggs (1940, p. 1007) on compression of marble at 150°C. in the presence of carbonated water. It is also advocated by Bain (1938, pp. 17, 18) who draws attention to the maximum compressibility and minimum solubility of calcite parallel to the *c* axis as factors possibly to be correlated with the orienting process.

In the case of Yule marble, the first two mechanisms listed above are rendered unlikely by the total lack of preferred orientation of lattice directions other than the *c* axis. However, the fabric data are compatible with orienting by crystallization. This could involve either selective growth of suitably oriented grains under the direct influence of compression normal to the foliation, or post-tectonic crystallization controlled by a pre-existing set of *s*-planes of undertermined origin.

Orientation of deformation lamellae. The visible deformation lamellae in Yule marble conform to a pattern which seems constant throughout the field of the investigated block. Its main features are as follows:

(1) Poles of lamellae, while necessarily concentrated in the vicinity of the direction of alinement of optic axes, also show a distinct though rather weak girdle, the axis of which (X in fig. 9) dips S. 30° E. at almost 35°.

(2) Where only two sets of lamellae are well developed in each grain (over 70° of the total number), the intersection [*e* : *e′*] of lamellae in any grain tends to be oriented parallel to the axis of the lamellae girdle (cf. fig. 20).

(3) There is equally strong orientation of the normal to [*e* : *e′*] within {0001}, around a corresponding direction in the lamellae girdle (cf. fig. 26).

(4) The glide line [*r* : *r′*] for twinning on $\{01\bar{1}2\}$ has random orientation (cf. fig. 18).

The distinctive lamellae pattern contrasts with, and cannot be directly related to the simple space-lattice orientation pattern discussed in the previous section. A reasonable explanation is that the lamellae pattern has been imprinted by late deformation upon a rock fabric which already possessed strong space-lattice orientation of the calcite grains. According to this view the lamellae developed wherever glide planes and directions were favorably oriented with regard to the deforming stresses.

It is instructive to compare the orientation of lamellae in Yule marble with corresponding data recorded by Felkel (1929) in her classic account of the fabric of calcareous tectonites from the Austrian Alps. This author, as well as measuring *c* axes and $\{01\bar{1}2\}$ lamellae, has recorded the orientations of such crystal directions as the glide line [*r* : *r′*] in $\{01\bar{1}2\}$, and the two bisectors of the angles between the two glide directions in such grains as show two sets of lamellae. The rocks in question include marbles with a single set of *s*-planes of slip, other rocks with several such *s*-planes intersecting in a B-axis, and yet others with a foliation developed by flattening. Some of Felkel's observations which may bear on the case of Yule marble are as follows:

(1) In rocks where there is a single set of *s*-surfaces (= *ab*) of slip, with corresponding strong concentration of lamellae parallel to *ab*, there is usually a distinct concentration of glide lines $[r : r']$ parallel to the *a* fabric axis; but these directions also spread from *a* into a continuous *ab* girdle (Felkel, 1929, figs. 10, 11). In B-tectonites in which the poles of lamellae are distributed through a B-girdle, orientation of glide lines $[r : r']$ is poor, the only regular feature being a broad irregular girdle more or less parallel to the foliation.

(2) Directions which nearly always show sharp orientation are the lines bisecting the angles between two $\{01\bar{1}2\}$ lamellae in any grain. The acute bisector, ***Rsyk***, is the same direction as is designated $\frac{\perp [e : e']}{(0001)}$ in this paper. The obtuse bisector ***Rsyg***, is a direction intermediate between the *c* axis and $[e : e']$ in any grain. There seem to be two alternative rules for orientation:—

(a) ***Rsyg*** is concentrated in the vicinity of the *a* fabric axis (Felkel, 1929, figs. 13, 29); ***Rsyk*** $= \frac{\perp [e : e']}{(0001)}$ is rather irregularly concentrated in the general vicinity of *b* (Felkel, 1929, figs 14, 28). The rocks in question are either S-tectonites with one of slip surfaces, or B-tectonites with intersecting sets of slip surfaces.

(b) ***Rsyk*** $= \frac{\perp [e : e']}{(0001)}$ is strongly concentrated parallel or close to the *a* fabric axis (Felkel, 1929, fig. 39). The example described is a rock with foliation of the flattened type with paired slip planes symmetrically inclined to the foliation. Alinement of ***Rsyk*** parallel to *a* and of ***Rsyg*** parallel to *b* (B) seems also to be indicated in the case of a B-tectonite believed by Felkel to be complicated by the presence of an incipient second B-axis (Felkel, 1929, 74-76, figs. 57, 58).

In the light of Felkel's data it is possible to identify tentatively the three fabric axes with which to correlate the deformation lamellae in Yule marble. The axis b' (X in fig. 9) is located approximately as the axis of the lamellae girdle, figure 9. The maximum concentration of edges $[e:e']$ in figure 20 coincides roughly with b' of the lamellae diagram; so that in Yule marble the second of Felkel's orientation rules, stated under paragraph (b) above, appears to hold good. The center of the broad maximum concentration area for $\frac{\perp [e:e']}{(0001)}$ defines the position of the a' fabric axis on this assumption. The $a'b'$ plane of the lamellae fabric thus approximates fairly closely to the foliation; but there is no correspondence between the b' axes for lamellae and the macroscopic lineation identified earlier as the b fabric axis.

Tectonic significance of the lamellae orientation pattern. At least some of the observed deformation lamellae must be of late origin. For example, many cases were observed where each lamella in one grain is matched by a similarly situated (but differently oriented) lamella continuing on across the immediately adjacent grain. Such lamellae can have formed only after both grains concerned had assumed their present orientation and outline.

The megascopic foliation is defined by parallel orientation of lensoid grains, and with it the space-lattice orientation of the calcite fabric is simply correlated. It is therefore likely that the observed space-lattice orientation, and the present orientation of grains according to external form developed simultaneously. The orientation pattern of the deformation lamellae has probably been influenced by the foliation; but it also shows clearly a girdle arrangement completely lacking in the space-lattice pattern. This condition, too, suggests a late origin for the lamellae. Moreover, the external shape of the elongated grains which dominate the foliation cannot be accounted for by movement on the observed lamellae within them. If it be assumed that lamellae are products of twin gliding, then in any lensoid grain the lamellae are too few and much too narrow to account for any perceptible elongation. If on the other hand the more effective mechanism of translation gliding is appealed to, the outline of a deformed grain should be notably offset wherever it is cut by a translation lamella. Significant offsetting or stepping of grain boundaries, though searched for carefully, was not found.

The fabric characters outlined above are all compatible with the hypothesis, now put forward, that the deformation lamellae represent the imprint of a late, relatively weak deformation insufficient to modify perceptibly a pattern of space-lattice orientation and a related foliation which had already developed as described in an earlier section. It is not necessary to assume that all visible lamellae belong to the one period of deformation. Indeed there are at least three ways in which lamellae may have formed; and in all probability lamellae of all three modes of origin may be present:—

(1) Lamellae may be expected to form to some extent during grinding of the thin sections. In the present case the number of such lamellae cannot be great, since so many sets of lamellae conform in some way to a pattern of preferred orientation which is common to all three sections.

(2) By analogy with the behavior of metallic aggregates and alloys (cf. Boas and Honeycombe, 1946, 1947), the internal stresses set up by differential expansion and contraction of crystals of calcite (a highly anistotropic material), during the long process of cooling and unloading that preceded exposure of the rock at the earth's surface, may well have been relieved to some extent by local translation or twin gliding. If the lamellae pattern in Yule marble is an expression of this kind of internal adjustment, all its features should be simply related to two directions which might conceivably have exerted an orienting influence, viz. the direction of pre-existing alinement of optic axes normal to the foliation, and the vertical direction of the force of gravity. No such simple relation has been observed.

(3) There remains the third possibility that late tectonic deformation is responsible for the appearance of at least such lamellae as conform to the orientation pattern here recorded.

Mechanism of gliding. Although deformation lamellae are conventionally described as "twin lamellae" in most text-books of petrography, in some rocks the twinned condition of many lamellae is by no means obvious. In Yule marble the great majority of observed lamellae belong to this doubtful category. When tilted parallel to the vertical axis of the microscope they appear as thin sharp lines resembling cleavage cracks rather than twin lamellae. If they are indeed twinned they cannot be distinguished from the host crystal by differences in optical behavior. There is an alternative possibility, namely, that many of these lamellae are due to translation gliding on $\{01\bar{1}2\}$, a mechanism experimentally demonstrated for single crystals of calcite by J. F. Bell (according to Fairbairn, 1942, p. 63). Both Felkel (1929, p. 79) and Sander (1930, p. 202) consider that pure translation, as well as twin gliding, on $\{01\bar{1}2\}$, is an important orienting mechanism.

Evidence strongly suggestive of translation gliding on $\{01\bar{1}2\}$ in the case of Yule marble is noted briefly as follows:—

(1) Definite twinning was recognized in only 10% of the 571 sets of lamellae measured in the course of this investigation.

(2) The glide line $[r : r']$ for twinning in $\{01\bar{1}2\}$ lacks any significant orientation in Yule marble (fig. 18). However this is perhaps mainly due to the weakness of the lamellae orientation pattern for the glide line $[r : r']$ seems likewise to lack preferred orientation even for cases of undoubted twinning (fig. 31). Moreover, the most likely glide line for translation in $\{01\bar{1}2\}$ is also $[r : r']$ (cf. Fairbairn, 1942, p. 63).

(3) Figures 29 and 30 show clearly that undoubted twinning is confined to grains having a range of space-lattice orientation which is narrowly limited when compared with that of the rock as a whole. Thus in figure 29, the optic axes of the host grains fall within the upper half of the diagram in only 25% of those cases where undoubted twinning occurs. A count of the plotted points (340) corresponding to figure 5 shows that the optic axes of all measured grains are evenly distributed, 50% in each half of the diagram. This strongly suggests that twinning, as contrasted with translation, occurs only in grains already having a favorable orientation of space-lattice. But it is also possible that all observed lamellae are twinned, and that twin gliding only proceeds far enough to develop relatively thick optically recognizable lamellae in grains with favorable orientation.

(4) Figure 32 shows that there is very strong preferred orientation of the optic axes within undoubted twin lamel-

lae. This diagram differs notably from those showing the orientation of optic axes in *all* lamellae on the assumption that these are indeed due to twinning (figs. 33, 34). The range of orientation in the latter is much broader than is the case for undoubted twin lamellae.

The case for translation gliding in Yule marble is still not definitely established. But the facts outlined are at least thoroughly compatible with the hypothesis that a high proportion, and perhaps even the great majority, of the deformation lamellae observed in this rock originated by translation gliding in $\{01\bar{1}2\}$. Strong preferred orientation of the directions $[e:e']$ and $\frac{\perp [e:e']}{(0001)}$ for paired lamellae suggests that the orientation of some potential glide line, as well as orientation of the plane $\{01\bar{1}2\}$, within the rock fabric has had some influence in determining whether or not gliding should take place on a given $\{01\bar{1}2\}$ plane in any given grain. By analogy with the cases of quartz (e.g. Fairbairn, 1941a; Ingerson and Tuttle, 1945) and olivine (Turner, 1942), preferred orientation of some glide line within the active glide plane is to be expected. Moreover, it is possible that translation on $\{01\bar{1}2\}$ may have been accompanied by translation on some other plane such as $\{0001\}$. Speculation beyond this point is not justified in the absence of any strict analysis of gliding possibilities in the calcite lattice, taking into account electrical properties of the component ions as well as their geometric arrangement in space.

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THE BASE OF THE CAMBRIAN IN THE SOUTHERN APPALACHIANS

PHILIP B. KING

PART II

OCOEE SERIES

General relations.—For a distance of 60 miles or more southwest of the Virginia-Tennessee boundary the Chilhowee group lies directly on the Cranberry and other granites equivalent to the injection complex of Virginia (fig. 1). If any intervening sedimentary or volcanic rocks once existed here, they were removed prior to deposition of the Chilhowee group.

Southwest of this point, from the French Broad River through southeast Tennessee into northern Georgia, the Chilhowee group is flanked on the southeast by rocks of the Ocoee series, which form the Great Smoky Mountains and other high ranges on the Tennessee-North Carolina border. The stratigraphy and structure of this area is much less clear than that farther northeast, and is undoubtedly more complex. Extensive overthrusts are present, and low to high-grade metamorphic features are nearly everywhere evident.

The Ocoee series was named by Safford (1856, pp. 151-152; 1869, pp. 183-198) for the Ocoee River, Polk County, Tennessee.[9] From Safford's time to the present, the age, origin, and relations of the Ocoee have been much debated and variously interpreted. The principal interpretations that have been made are indicated in table 2.

Some of these interpretations have been outmoded by the acquisition of contrary evidence. Some represent successive interpretations by the same geologists—for example, interpretations 2, 3, and 5 are by Keith, and interpretations 6, 7, and 8 are by Stose and Stose. In this paper, it does not seem necessary to consider the evidence for or against all the interpretations. The only one that will be considered at length is the last one by Keith (interpretation 5, table 2), after which an alternative of the writer's (interpretation 10) will be proposed.

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[9] Problems relating to usage of the term Ocoee series are discussed under the heading "Notes on terminology" at the end of this paper.

TABLE 2

Interpretations that have been made regarding age and relations of the Ocoee series

	Date	Author and reference	Age	Stratigraphic relations
1	1869	Safford, Geology of Tennessee	Potsdam group of Lower Silurian; also Eozoic	Lies stratigraphically beneath Chilhowee group; relations gradational.
2	1889 1891	Keith, unpublished work Willis, in Walcott, U. S. Geol Survey Bull 81, pp. 299-300	Ordovician or younger	Unconformable on Ordovician rocks of Appalachian Valley; clastic southeastern facies of upper Ordovician and younger Paleozoic rocks
3	1895	Keith, Knoxville folio (number 16)	Age unknown	Ample evidence which separates it from the Cambrian series, though not sufficient to fix its age.
4	1895	Hayes, Cleveland folio (number 20)	Age unknown, probably Algonkian	Bears all the marks of extreme age so that it is best to consider it Algonkian until satisfactory evidence to contrary is found.
5	1904 1907	Keith, Asheville and Nantahala folios (numbers 116 and 143)	Lower Cambrian	In part lateral southeastern equivalent of Chilhowee group, in part older
6	1932 1932 1936	Jonas, AM JOUR SCI., 5th ser, vol. 24, pp 240-241. Geologic map of United States. Crickmay, Geol Soc. Amer. Bull, vol. 47, p. 1391.	Partly Lower Cambrian, partly pre-Cambrian	Ocoee belt is divided longitudinally by Cartersville overthrust, masked by a zone of retrogressive metamorphism, which brings together two units of similar lithology but of very different age.
7	1944	Stose and Stose, AM. JOUR. SCI, vol 242, p. 416.	Pre-Cambrian	Nearly everywhere separated from Chilhowee group to northwest by overthrust fault; probably unconformable below Chilhowee
8	1946	Stose and Stose, Geol. Soc Amer. Bull, vol. 57, p. 1233	Partly Ordovician, partly pre-Cambrian	Shales, limestones and sandstones along northwest side of Ocoee belt are Ordovician, and separated by overthrust fault from remainder of Ocoee series.
9	1948	Lombard, Amer. Assoc Petrol. Geol Bull., vol 32, p 720.	Possibly Paleozoic	May be an allochthonous southeastern clastic facies of some or much of Paleozoic.
10	1949	King, this paper.	Later pre-Cambrian	Lies stratigraphically beneath Chilhowee group, but formed during an earlier cycle of sedimentation; possibly unconformable below Chilhowee, at least locally.

Interpretation by Keith.—Some of the interpretations indicated in table 2 imply a complete or nearly complete separation of the Ocoee series by faults from the Chilhowee group and other rocks of Paleozoic age to the northwest. Lombard (1948, p. 720; interpretation 9, table 2) thus suggests that the series may be an allochthonous southeastern facies of a considerable part of the Paleozoic. Stose and Stose (1944, p. 416; interpretation 7), while considering the series allochthonous, believed it to be of late pre-Cambrian age.

Keith, on the basis of his comprehensive field studies of the Ocoee series, concluded on the other hand, that the series was closely related to the Chilhowee group, and that the separation of the two by faulting was local and minor. This interpretation has been confirmed by the writer and his colleagues, who likewise conclude that the Ocoee series and Chilhowee group are not only contiguous but closely related units.

Rocks here regarded as a part of the Ocoee series were divided by Keith into a number of formations, but as the terminology varies from one folio report to another, the record is difficult for the uninitiated to follow. Terminology used in the oldest report, the Knoxville folio (Keith, 1895, pp. 2-3) represents a preliminary attempt at subdivision which was subsequently found unusable. The terminology used in the later folios, such as the Asheville (1904, pp. 4-6) and the Nantahala (1907, pp. 3-5) reflects Keith's more mature judgment, and applies to units that can be recognized in the field. It was also used by Keith in his unpublished work on the Mount Guyot, Cowee, and Murphy quadrangles. Some of the later terminology and interpretation was based on Keith's unpublished work in these three quadrangles, so that the evidence for it is not generally available to geologists.

Table 3 lists Keith's later classification and the earlier classification in the Knoxville folio. Correlations given between the two columns are only approximate.

According to Keith's later work, in the area northeast of the Great Smoky Mountains (Mount Guyot and Asheville quadrangles) the Cochran conglomerate of the Chilhowee group lies directly on the Hiwassee slate, which is well down in the column on table 3. In the Great Smoky Mountains, the Great Smoky conglomerate and the Nantahala slate are interpreted as overlying the Hiwassee slate. There is thus an excess

TABLE 3

Interpretation of sequence in Ocoee series according to Arthur Keith

	Keith, Asheville folio (1904), Nantahala folio (1907), and Mount Guyot sheet (unpublished)	Keith, Knoxville folio (1895)
	Overlying formations of the Murphy marble belt	(Top eroded)
Ocoee series	Nantahala slate	Clingmans conglomerate
		Hazel slate
	Great Smoky conglomerate	Thunderhead conglomerate
		Cades conglomerate
	Hiwassee slate	Pigeon slate
		Citico conglomerate
		Wilhite slate
	Snowbird formation	(Base concealed)
	Max Patch and Cranberry granites; Carolina gneiss	

of younger Ocoee-type rocks in the Great Smoky Mountains, as compared with the area to the northeast. This was explained by Keith by assuming that the Great Smoky conglomerate and Nantahala slate were equivalent to the Cochran conglomerate and Nichols shale of the Chilhowee group. Keith therefore believed that the Ocoee series was in part equivalent to the Chilhowee group, and was in part older. Throughout his later reports he therefore classed the formations belonging to the Ocoee series as of Lower Cambrian age.

This interpretation, while superficially plausible, fails to account for many features. The Great Smoky conglomerate is more than twice as thick as the Cochran conglomerate; moreover, it is of graywacke facies, whereas the Cochran is of arkosic and quartzose sandstone facies. These differences were supposed by Keith to result from lateral variations in the deposit. To the writer, they appear so great that he does not believe the two formations could have been laid down in the same basin of deposition; hence it is unlikely that they are of the same age. Moreover, no such change from one facies to another has ever been traced out along the outcrop. In northeastern Tennessee the Unicoi formation (Cochran equivalent) crops out in successive outcrop belts, to a breadth of 35 miles across the strike, and the distance before deformation was probably twice as great. Variations in the Unicoi exist between

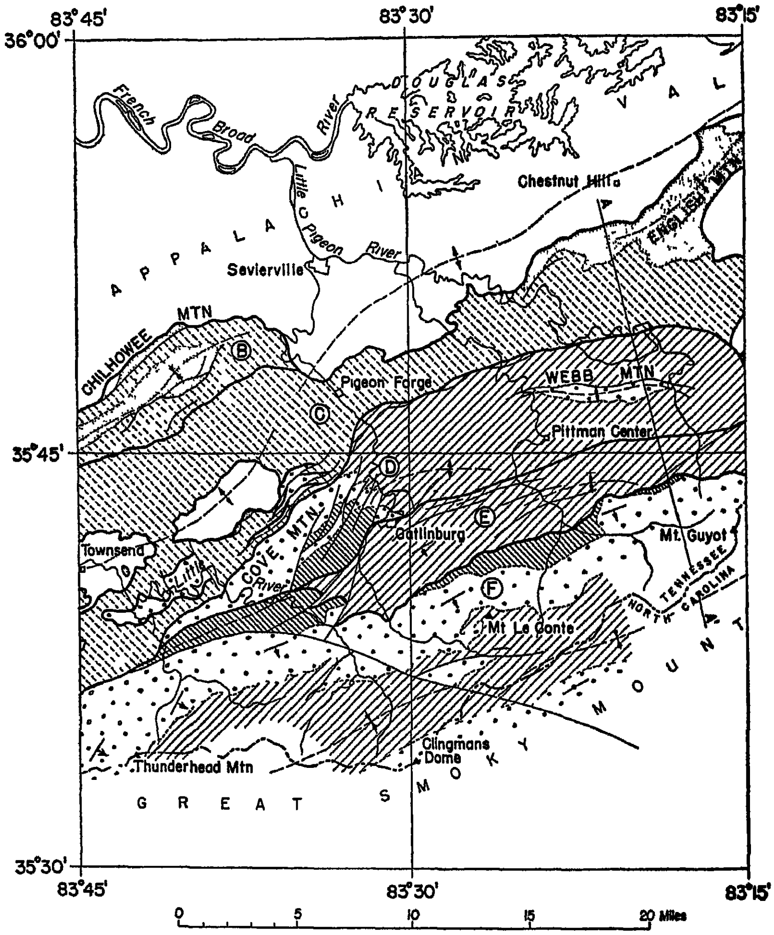
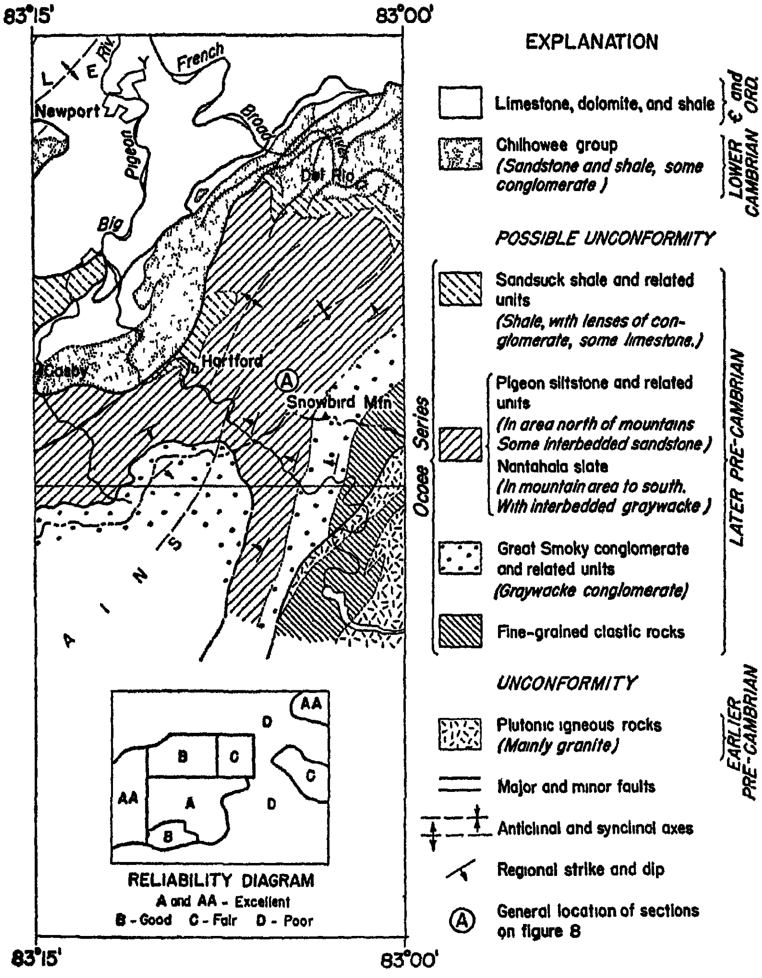


Figure 7. Preliminary geologic map of northeast part of Great Smoky Mountains and adjacent foothills to north, based on detailed and reconnaissance surveys by J. B. Hadley, H. W. Ferguson, and the writer.

the different outcrop belts, but the rocks remain recognizably Unicoi, and never assume a Great Smoky aspect. Moreover, it is highly unlikely that the beds beneath the Cochran and Great Smoky are the same, although each were termed Hiwassee by Keith. The formation beneath the Cochran is an argillaceous shale, with beds of conglomerate and limestone; the formation beneath the Great Smoky is siltstone, with beds of fine-grained graywacke sandstone.


83°15'
83°00'
Newport
French
Broad
Pigeon
Big
Del Rio
Cosby
Hartford
Snowbird Mtn
Ⓐ
EXPLANATION
Limestone, dolomite, and shale
€ and ORD.
Chilhowee group
(Sandstone and shale, some conglomerate)
LOWER CAMBRIAN
POSSIBLE UNCONFORMITY
Sandsuck shale and related units
(Shale, with lenses of conglomerate, some limestone.)
Pigeon siltstone and related units
(In area north of mountains Some interbedded sandstone)
Nantahala slate
(In mountain area to south. With interbedded graywacke)
Great Smoky conglomerate and related units
(Graywacke conglomerate)
Fine-grained clastic rocks
Ocoee Series
LATER PRE-CAMBRIAN
UNCONFORMITY
Plutonic igneous rocks
(Mainly granite)
EARLIER PRE-CAMBRIAN
Major and minor faults
Anticlinal and synclinal axes
Regional strike and dip
General location of sections on figure 8
RELIABILITY DIAGRAM
A and AA - Excellent
B - Good C - Fair D - Poor
AA
B
C
A
D

Regional interpretations of the stratigraphy and structure are those of the writer. The geology in areas marked AA on the "reliability diagram" is published with permission of the Tennessee Division of Geology.

Tentative interpretation by the writer.—Work on the Ocoee series in Great Smoky Mountains National Park is now in progress by parties of the U. S. Geological Survey and Tennessee Division of Geology. On the basis of detailed and reconnaissance work by these parties, the writer has arrived at a tentative interpretation of the sequence which differs from that of Keith (interpretation 10, table 2). This is briefly outlined here, and is illustrated by a map (fig. 7) and by stratigraphic sections (fig. 8). Final interpretation of the Ocoee series must await further field work by the present parties, and this work will no doubt result in more or less modification of the interpretation here given.

The beds beneath the Cochran conglomerate northeast of the Great Smoky Mountains (the so-called Snowbird formation and Hiwassee slate) are believed to be equivalent to the entire Ocoee series farther southwest, rather than its lower part alone. To the northeast these beds are relatively thin, but to the southwest, near the Big Pigeon River, they attain a thickness of about 17,500 feet (section A, fig. 8). The shales at the top, the so-called Hiwassee, are similar in character to the Sandsuck shale, a formation which underlies the Cochran conglomerate of the Chilhowee group in its type area on Chilhowee Mountain (Section B, fig. 8). The Sandsuck shale is believed to be the top formation of the Ocoee series, and to be younger than any rocks in the Great Smoky Mountains to the southeast. The strata which were correlated with it by Keith, and which were supposed to underlie the Great Smoky conglomerate of the mountains, are actually much older (section E, fig. 8). The formations in the Great Smoky Mountains are believed to be equivalent to parts of the so-called Snowbird formation near the Big Pigeon River, and to have been raised successively higher toward the southeast by thrust faults (structure section A, fig. 9). For example the Great Smoky conglomerate (section F, fig. 8) is probably equivalent to conglomeratic beds in the lower part of the Snowbird formation on the Big Pigeon River. According to this tentative interpretation, the sequence of the Ocoee series is therefore as shown in table 4.

According to this proposed interpretation, the Ocoee series underlies the Chilhowee group, is entirely older than the Chilhowee group, and no part of either unit is equivalent to any

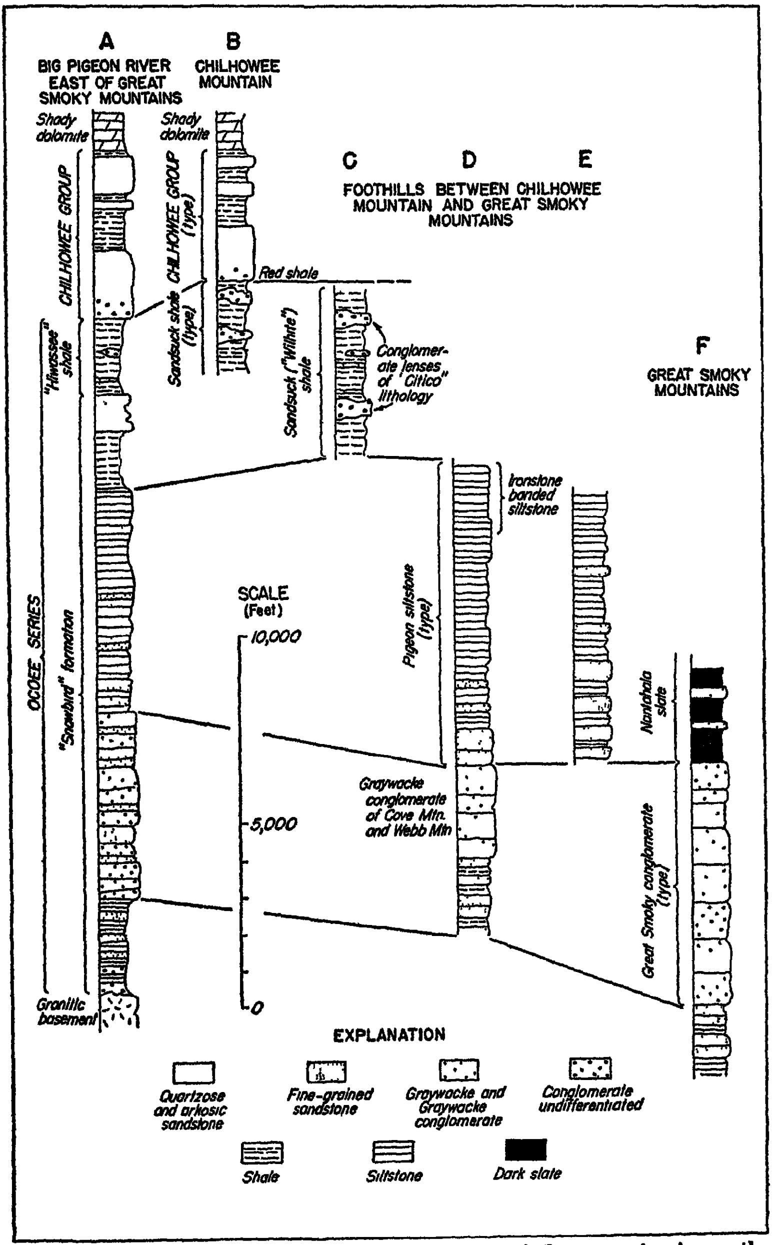
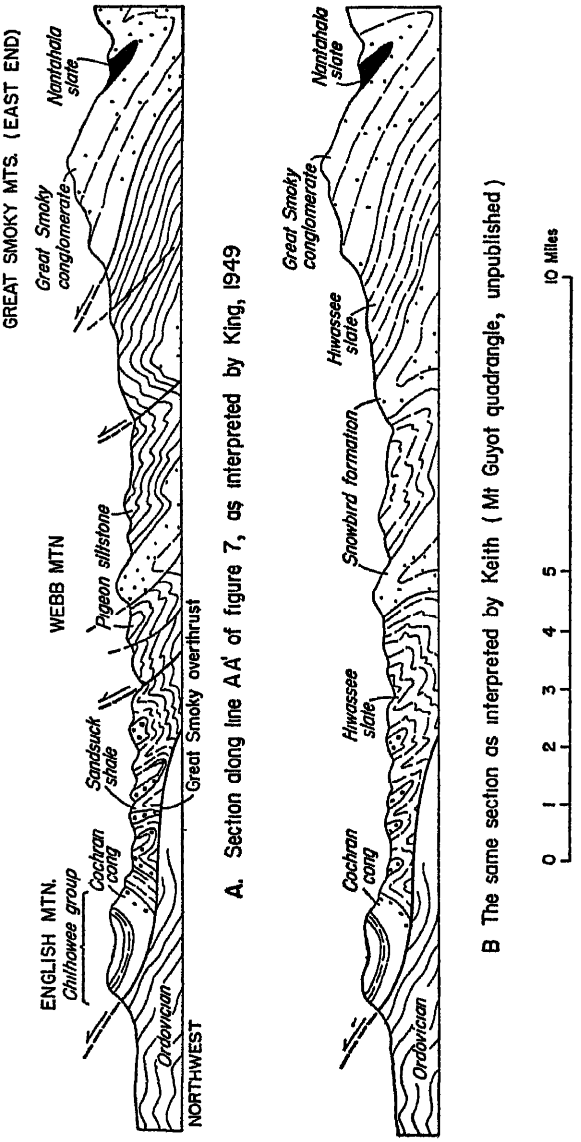

A
BIG PIGEON RIVER EAST OF GREAT SMOKY MOUNTAINS
B
CHILHOWEE MOUNTAIN
Shady dolomite
Shady dolomite
CHILHOWEE GROUP
CHILHOWEE GROUP (type)
"Hiwassee" shale
Sandsuck shale (type)
Red shale
C D E
FOOTHILLS BETWEEN CHILHOWEE MOUNTAIN AND GREAT SMOKY MOUNTAINS
Sandsuck ("Wilhite") shale
Conglomerate lenses of "Citico" lithology
F
GREAT SMOKY MOUNTAINS
Ironstone banded siltstone
OCOEE SERIES
"Snowbird" formation
SCALE (Feet)
10,000
5,000
0
Pigeon siltstone (type)
Nantahala slate
Graywacke conglomerate of Cove Mtn. and Webb Mtn
Great Smoky conglomerate (type)
Granitic basement
EXPLANATION
Quartzose and arkosic sandstone
Fine-grained sandstone
Graywacke and Graywacke conglomerate
Conglomerate undifferentiated
Shale
Siltstone
Dark slate

Figure 8 Preliminary stratigraphic sections of Ocoee series in northeast part of Great Smoky Mountains and adjacent foothills to north, to illustrate tentative interpretation by the writer. Location of sections is shown in figure 7. Sections A and B are based mainly on work of H. W. Ferguson, and are published with permission of the Tennessee Division of Geology.

part of the other. The Ocoee is a distinct series of great thickness, characterized by its own sequence and its own lithologic features.

Basement on which series was deposited.—The Ocoee series contains great quantities of material derived from preexisting plutonic rocks, including much "blue" quartz like that in the



A. Section along line AA' of figure 7, as interpreted by King, 1949

B The same section as interpreted by Keith (Mt Guyot quadrangle, unpublished)

Figure 9. Structure sections along line A A' of geologic map, figure 7, showing interpretations by the writer, and by Keith.

TABLE 4

Tentative interpretation of sequence in Ocoee series in northeast part of Great Smoky Mountains, as proposed by P. B. King

Unit	Representative formations included in unit	Character	Approximate thickness (in feet)
	Chilhowee group at top of sequence (Cochran conglomerate is lowest formation).		
4	Sandsuck shale	Argillaceous shale, with interbedded layers and lenses of limestone and of conglomerate (latter is the Citico conglomerate of Keith's early reports).	4,000
3	Pigeon siltstone (north of mountains) Nantahala slate (in mountain area)	North of mountains, laminated siltstone and interbedded fine-grained graywacke sandstone In mountain area, dark gray or black slate, with interbedded coarse graywacke.	8,000
2	Great Smoky conglomerate	Medium to coarse graywacke sandstone and conglomerate, composed of quartz and feldspar grains, in thick beds, with graded bedding	6,000
1		Fine-grained sandstone, with interbedded siltstone and slate	2,500
	Base not exposed in Great Smoky Mountains; equivalent beds near Big Pigeon River rest on Max Patch and Cranberry granites.		

Max Patch and Cranberry granites of the Asheville quadrangle, North Carolina (Keith, 1904, pp. 3-4). In the Asheville quadrangle, Keith mapped the Snowbird formation of the Ocoee series as lying unconformably on these granites. Reconnaissance observations by the writer and his colleagues suggest that this relation is probably correct, although in places there are structural complications at the contact.

On Cohutta Mountain, east of Chatsworth, Murray County, Georgia, Furcron and Teague (1947, pp. 19-21, pl. 1) have found that older granites and gneisses are exposed in a narrow strip along the northwest margin of the belt of outcrop of the Ocoee series, and they believe that the series lies unconformably on the granites and gneisses (fig. 10).

In other places, the basal relations of the Ocoee series are less certain. In the Nantahala quadrangle, North Carolina (Keith, 1907) and the Ellijay quadrangle, Georgia (La Forge and Phalen, 1913) the Ocoee series is mapped as lying unconformably on the Carolina gneiss, classed by Keith as of Archean age. Crickmay (1936, pp. 1380-1381) and others have expressed doubt as to this interpretation, and have pointed out that the more highly metamorphosed phases of the Ocoee closely resemble the Carolina gneiss. This suggests that the one is perhaps merely a greatly metamorphosed facies of the other. In western North Carolina, in the Mount Guyot and Cowee quadrangles, Keith (unpublished reports) mapped granites which were supposed to be pre-Ocoee, and against which the Ocoee series was supposed to overlap unconformably. These granites are definitely intrusive into the Ocoee (J. B. Hadley, personal communication, 1948).

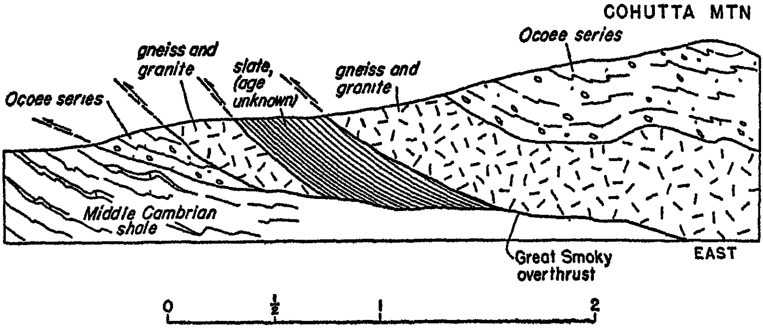


Figure 10. Section of Cohutta Mountain, east of Chatsworth, Murray County, Georgia, showing relation of Ocoee series to older granites and gneisses. After Furcron and Teague, 1947.

Conditions of deposition.—The Ocoee series consists entirely of sedimentary rocks all of which are clastic, except for rare thin beds of limestone; no volcanic rocks are known. The Ocoee forms a tremendously thick mass. Keith reports thickness of 7,500 to more than 10,000 feet in the Knoxville and Nantahala quadrangles (1895, 1907), but these are certainly underestimates. As indicated by table 4, the thickness may exceed 20,000 feet, and the series thus has a mass nearly equal to that of the Paleozoic rocks in the Appalachian Valley to the northwest.

The sedimentary rocks of the Ocoee series form a distinctive suite, quite unlike those of other nearby clastic units, such as the Chilhowee group. Some of the differences between the Ocoee series and the rocks of the Chilhowee group have been listed by Stose and Stose (1944, pp. 404-405). The lithologic, petrographic, and chemical character of the rocks of the Ocoee series have been described by Moneymaker (1938). Conditions of deposition of the Ocoee series have been outlined by Barrell (1925, pp. 15-20), although his conclusions must be modified to the extent that he accepted Keith's interpretation of the intergradation of the Ocoee series and the Chilhowee group.

The rocks of the lower part of the Ocoee series (unit 2, table 4) are of graywacke facies[10] and strongly resemble the graywackes of the Archean of the Canadian shield as described by Pettijohn (1943, pp. 941-960). Various textures occur, ranging from fine and medium-grained sandstones (in unit 3, table 4) to fine-grained conglomerates (unit 2), but there are no quartzose or arkosic sandstones like those in the Chilhowee group. The graywackes contain many grains of feldspar and quartz, but the matrix is argillaceous material, now altered to sericite and chlorite. The conglomerates contain few pebbles larger than half an inch in diameter, and practically all are single mineral grains or crystals, as though they had been derived from wholesale granular disintegration of the source rocks. Barrell (1925, p. 19) has noted that Ocoee rocks are characteristically gray, bluish gray, or even black-colored, implying rapid sedimentation and lack of thorough oxidation before burial.

In the graywackes of the Ocoee series, as in others, graded bedding is a striking feature, which suggests that the sediments were deposited rapidly, without sorting except that which was produced by settling in water. This is manifested by cyclical arrangement of the coarse and fine deposits, each cycle beginning with granular or pebbly deposits, followed by sandy and then silty or slaty deposits, the whole cycle being represented by beds a foot to more than 25 feet thick, and endlessly repeated through thousands of feet of section. In many places the slaty or silty beds at the tops of the layers are broken up and reincorporated as chips or slabs in the basal deposits of the succeeding unit. Ripple marks and cross bedding are rare. Pettijohn notes that these two features seem to be caused by different conditions of sedimentation from graded bedding, and that they are seldom found together.

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[10] *For explanation of rock names used, see "Notes on terminology" at end of paper.*

The succeeding beds (units 3, table 4) north of the mountains are largely siltstone, which forms a very thick and uniform body. This rock type, intermediate in texture between a sandstone and a shale, is relatively uncommon in nearby geological sequences, such as that of the Paleozoic rocks of the Appalachian Valley, where less rapid sedimentation permitted more thorough sorting of the constituents. A change to more normal conditions of sedimentation takes place in the final unit of the Ocoee series (unit 4, table 4), which contains beds of limestone, and whose shales and conglomerates are not unlike those in the succeeding Chilhowee group. The conglomerates of the final unit are not of graywacke facies, as are those in the lower part of the series. They were probably laid down under different conditions and derived from another source.

Whether the Ocoee series is unconformable below the Chilhowee group is as yet undetermined. Apparently no formations of the Ocoee series were cut out by erosion before Chilhowee time, and the Sandsuck shale (unit 4) forms the top at all places. Some evidence has been found by H. W. Ferguson (personal communication, 1949) that the Sandsuck beds were at least moderately eroded before Chilhowee time, but the significance of this remains to be evaluated.

Regardless of whether or not the Ocoee series and Chilhowee group are unconformable, the two units were laid down under different conditions, and during separate cycles of sedimentation. The sediments of the Ocoee series were deposited rapidly, in a region of high relief and great tectonic instability, in a relatively narrow, rapidly subsiding trough—a eugeosyncline. The sequence begins with coarse clastic deposits (units 1 and 2, table 4), whose source rocks in surrounding lands wasted by granular disintegration, and which were deposited before complete oxidation and with no sorting except that produced by settling in water. These are succeeded by silty sediments, laid down to a vast thickness, and finally by argillaceous sediments, which contain beds of limestone. Chilhowee time again begins with the deposition of coarse clastic deposits, but these are of arkosic and quartzose sandstone facies, and are the initial deposits of the Appalachian miogeosyncline of Paleozoic time.

GEOLOGIC HISTORY

This survey of the stratigraphy of the older Cambrian and the pre-Cambrian rocks of the southern Appalachians suggests the following possible events in the geologic history:

1) Prolonged erosion of a plutonic terrane, composed of such rocks as the Max Patch and Cranberry granites in Tennessee, and the injection complex in Virginia.

2) Deposition in restricted basins of sedimentary and volcanic rocks, including (a) the Catoctin greenstone and underlying sedimentary rocks, (b) the volcanics of the Mount Rogers area, and (c) the Ocoee series. All were formed under conditions of considerable crustal instability, manifested in the first two units by volcanic action and in the third by the development of a rapidly subsiding trough bordered by lands of high relief, a eugeosyncline. The origin and history of this crustal instability does not concern us directly here, except to say that it took place before the depositions of the Chilhowee group.

3) Remnants of the deposits of these basins are now widely separated from each other, the Chilhowee group in the intervening areas lying directly on the older plutonic rocks. This separation results either from removal of the original connecting strata from between the basins, from original isolation of the basins, or from both. No evidence exists to indicate whether deposition in the basins took place at the same or at different times. Very possibly the rocks of the different basins were not contemporaneous, yet they cannot be far apart in age.

4) Tilting of the margins of the basins during and at the close of the period of deposition, and erosion and truncation of the edges of the tilted beds. Such features have been seen by the writer at the edge of the Catoctin greenstone in the Elkton area and at the edge of the volcanics of the Mount Rogers area. An unconformity may also separate the Ocoee series from the Chilhowee group.

5) Deposition of the Chilhowee group as the initial deposit of the Appalachian miogeosyncline. The earliest deposits of the Chilhowee seem to be those which form the lower part of the Unicoi formation of northeast Tennessee and they may pass out by overlap in adjoining areas. In northern Virginia, at least, the time of their deposition is represented by a hiatus.

6) At intervals during early Chilhowee time there was renewal of volcanic activity. In northeast Tennessee, the main volcanic episode took place some time after the beginning of deposition of the Unicoi formation; consequently this volcanic episode took place at a much later time than the eruption of the volcanics of the Mount Rogers area. It is suggested that the volcanics of the Loudoun formation of northern Virginia are similarly much younger than those of the underlying Catoctin.

7) The first indications of life appear in upper Chilhowee rocks, the oldest being worm tubes of the *Scolithus* type. Trilobites and brachiopods do not appear until near the top of the Chilhowee group. The occurrence of these evidences of life in rocks of the Chilhowee group does not seem to be the result of abrupt evolutionary development, but may result from migration of marine life into a gradually improving environment.

8) Deposition of succeeding fossiliferous formations of Lower Cambrian age, beginning with the Shady and Tomstown dolomites.

CONCLUSIONS

The facts and interpretations just summarized suggest the following tentative conclusions as to the position of the base of the Cambrian in the southern Appalachians:

Proposals by the Cambrian subcommittee (Howell and others, 1944), and by Wheeler (1947) and Snyder (1947), to restrict the base of the Cambrian to the lowest horizon or formation that contains diagnostic Cambrian fossils have the virtue of precision, but seem unnecessarily restricted and arbitrary. Use of the base of the lowest formation that contains Cambrian fossils as a boundary is impracticable when one considers that formation boundaries in conformable sequences of beds are merely arbitrary units of mapping. The Erwin quartzite, for example, as it is traced southwestward, is divided into three formations. If the base of the Erwin is termed the base of the Cambrian, the boundary must be jogged upward where the division takes place, either to the base of the Murray shale or the base of the Hesse quartzite, depending on which contains the lowest fossil horizon. The use of the lowest horizon to contain diagnostic fossils has more to recommend it, but this depends largely on accidents of collecting. Thus, at one locality fossils are reported from the Murray shale; elsewhere the lowest fossil horizon is found at the top of the overlying Hesse or its equivalents. Possibly the most definitive boundary between Cambrian and pre-Cambrian would be a fossil zone with a distinctive assemblage of fossils, even if occasional fossils should occur beneath.

Such a definition might be highly satisfying for precise paleontological zonation, yet would be unsatisfying to the general geologist and stratigrapher who must deal with such rocks in the field. Except for the fact that the highest part of the Chilhowee group is fossiliferous and the remainder unfossiliferous, it is a closely knit sequence of deposits laid down in conformable succession, with none of its parts greatly different in age from the other parts. Under the circumstances it would seem to be impracticable to term one part Cambrian and another pre-Cambrian. The supposed pre-Cambrian part is the logical beginning of the supposed Cambrian part.

On the other hand, the writer can agree with Snyder (1947, p. 152) as to the undesirability of the practice of extending the Cambrian downward through great thicknesses of unfossiliferous strata until a major unconformity is reached. Major unconformities, with subaerial or subaqueous deposits lying on plutonic rocks, such as those described by Bloomer and Bloomer (1947, p. 95) below the Oronoco formation in Virginia, and by Furcron and Teague (1947, p. 20) below the Ocoee series in Georgia, are undoubtedly significant mileposts in geologic history. They do not, however, necessarily mark the base of the Cambrian. Similar major unconformities between strata recognized by all to be pre-Cambrian are well known in the Lake Superior region, the Grand Canyon district, and elsewhere.

Conditions could conceivably exist where the fossiliferous Cambrian was separated from the next major unconformity beneath by a vast thickness of unfossiliferous sediments, with a gradual transition downward from distinctly Cambrian-like rocks to distinctly non-Cambrian-like rocks. Such a condition in southern Nevada is described by Wheeler (1947, pp. 155-157). In the writer's opinion, these conditions do not exist in the southern Appalachians. Cambrian-like rocks are underlain by non-Cambrian-like rocks, but the first are separated from the second by changes in sedimentation, and in part at least by unconformities that involve tilting, rearrangement of the basins of deposition, and a moderate time hiatus. The extent of the basins of deposition and the environments of sedimentation were very different in each.

The writer concludes that the base of the Cambrian in the southern Appalachians can logically be placed at the base of the Chilhowee group, and that this group and its component formations forms the lowest unit of the Lower Cambrian series. He believes that the Catoctin greenstone and underlying sedimentary rocks, the volcanics of the Mount Rogers area, and the Ocoee series are probably of late pre-Cambrian age, although they may not necessarily be entirely contemporaneous.

NOTES ON TERMINOLOGY

Stratigraphic names.—The *Catoctin greenstone* was named by Keith (1894, p. 306), who called it the Catoctin schist. Most of the formation is not true schist and much of it is relatively massive. It has been termed a metabasalt in many geologic reports, but Bloomer and Bloomer (1947, p. 97) term it a propylite and interpret it as an altered andesite. As opinions differ regarding its precise petrographic classification, and as it may vary in composition from place to place, it seems most appropriate to use for it the general rock term greenstone, which was employed on the geologic map of Virginia (1928).

The *Ocoee series* was named by Safford (1856, pp. 151-152; 1869, pp. 183-198). Keith, in his subsequent folio work in the area did not use the term and preferred to speak of the sequence in terms of the component formations. However, the term Ocoee is desirable, as the whole unit is more distinctive than its component parts, and as it differs greatly in its gross features from adjacent rock units. In this paper, the Ocoee is classed as a provincial series, probably belonging to the late pre-Cambrian. The Ocoee series is thus comparable to such other pre-Cambrian provincial series as the Keweenawan series, the Belt series, and the Grand Canyon series. Like them, it is a thick body of sedimentary rocks which can be subdivided locally into a number of formations.

The Ocoee series may be defined as comprising Keith's Snowbird formation, Hiwassee slate, Great Smoky conglomerate, Nantahala slate, and Sandsuck shale. As indicated on table 4, the writer believes that the sequence as interpreted by

Keith will require considerable rearrangement. Also, some of Keith's terms will need to be abandoned or more precisely defined, and some of the units will need to be subdivided.

The *Snowbird formation* at its type locality is divisible into a number of well-marked units which are probably equivalent to named units in the Great Smoky Mountains section, and it should be subdivided on this basis. The name might be useful, however, for the thinned equivalents of these formations toward the northeast.

The *Hiwassee slate* was named for the Hiwassee River, Polk County, Tennessee, where it forms a belt of outcrop many miles wide. In the Great Smoky Mountains the term should be abandoned, as it was here applied to diverse units in the Ocoee series, whose relations to beds in the type area are unknown.

The type area of the *Great Smoky conglomerate*, according to Wilmarth (1938, p. 865), is in the Great Smoky Mountains of the Knoxville quadrangle; probably it also includes outcrops of the formation in the Mount Guyot quadrangle to the east. It is here capable of precise delimitation, and should be restricted to the unit as defined in this area and its equivalents. Farther south, as in the Nantahala and Ellijay quadrangles, the term has come to be used as a general expression for rocks of Ocoee type. This usage is improper, and the term Ocoee series should be substituted.

The *Nantahala slate* has its type area in the Nantahala quadrangle. This area has not been studied by the present party of the U. S. Geological Survey, but the rocks exposed there appear to be the same as those to which the term is applied in the Great Smoky Mountains.

The *Sandsuck shale* was named for Sandsuck Branch, on Chilhowee Mountain, in the Knoxville quadrangle. Here the formation is directly overlain by the Chilhowee group; hence, the term can be more precisely defined than the terms Wilhite slate and Hiwassee slate which Keith elsewhere used for equivalent beds.

In addition to these, the writer wishes to employ the term *Pigeon siltstone*, deriving the name from the same type area as the Pigeon slate of Keith's early work. According to Wilmarth (1938, p. 1658), this type area is the West Fork of the Little Pigeon River in the Knoxville quadrangle. Keith

(1895, p. 2) states that the Pigeon is extensively exposed in the drainage area of the Little Pigeon River, thus implying that it is also present along the Middle and East Forks of this stream in the Mount Guyot quadrangle. The Pigeon siltstone of present usage is widely exposed in the drainage areas of all three streams (fig. 8).

The formations of the Murphy marble belt (Tusquitee quartzite, Brasstown schist, Valleytown formation, Murphy marble, Andrews schist, and Nottely quartzite), which lie within the outcrop area of the Ocoee series (fig. 1), have been considered by some authors to be a part of the Ocoee series, but not by Safford. Their age and relations are uncertain, and they are not included in the present definition of the Ocoee series. In Safford's descriptions of outcrops in east Tennessee, he assigned to the Ocoee series in many places rocks which are now classed as parts of the Cochran and Unicoi formations. These are here excluded from the Ocoee and placed in the Chilhowee group.

The *Chilhowee group*, like the Ocoee series, was named by Safford (1856, pp. 152-153; 1869, pp. 198-203). It was not widely used by Keith in his folio work for he preferred, as in the case of the Ocoee series, to refer to the Chilhowee rocks in terms of their component formations. The need for a group name is indicated by the wide use of such informal terms as "basal quartzites" (Butts, 1933, p. 2), "Lower Cambrian quartzites and slates" (Geologic Map of United States, 1933), and "basal clastic group" (King and others, 1944, p. 27). The term Chilhowee group is appropriate for this purpose and has been so used by Butts (1940, pp. 26-27) and other authors.

According to Wilmarth (1938, p. 430) "the U. S. Geological Survey draws the base of the Chilhowee group at the base of the Cochran conglomerate and the top at the top of the Hesse sandstone." This usage corresponds with the usage of Butts and is followed in the present paper.

Unfortunately, Safford did not define exactly the base of the Chilhowee. In his description of Chilhowee Mountain (1869, p. 190) he states that strata of the Ocoee series are there exposed beneath strata of the Chilhowee group, which indicates that he placed part or all of what are now termed the Sandsuck shale and Cochran conglomerate in the Ocoee series. Elsewhere in east Tennessee, he generally excluded beds now classed as part of the Cochran conglomerate and Unicoi formation from the Chilhowee, although he was not entirely consistent. A similar usage has been followed by the Tennessee Division of Geology in its geologic map of Tennessee, where beds here termed the Ocoee series and the Chilhowee group have been divided into an upper, or "Chilhowee" division, and a lower, or "Unicoi" division. The former included the Hesse quartzite, Murray shale, Nebo quartzite, and Nichols shale, and their correlatives. The latter included the Cochran conglomerate, the Unicoi formation, and all the Ocoee series below the Nantahala slate. According to the interpretations of the present writer, grouping of the Cochran and Unicoi with the Ocoee series is unnatural.

Lithologic titles have been given to many formations of the Chilhowee group—such as the Nebo, Hesse, Erwin, and Antietam quartzites, and the Nichols, Murray, Hampton, and Harpers shales. These lithologic titles are not useful in detailed work because the supposed lithologic units actually contain many other types of rocks. In the northeast Tennessee report (King and others, 1944, pp. 27-28) the lithologic titles were abandoned and the units were termed formations. In regional work the lithologic titles have some value as they express broad changes in sedimentation, and they are therefore employed in this paper. The term quartzite is used in this paper where it applies to a lithologic title of a formation; it corresponds to indurated quartzose sandstone and arkosic sandstone as used elsewhere in the paper.

Rock names.—The writer wishes, for purposes of this paper, to distinguish three types of medium-grained clastic deposits, which are termed *quartzose sandstone, arkosic sandstone and conglomerate*, and *graywacke sandstone and conglomerate.* The three types are readily distinguishable megascopically and in the field, and work that has been done so far by Mr. Hadley suggests that they may show equally well marked differences under the microscope. The first two types occur in the Chilhowee group and are more or less gradational, although the end members are distinct. The third type occurs in the Ocoee series and contrasts strongly with the other two. Published terminology of medium-grained rocks shows considerable variation from author to author,[11] and the terms as here used differ from some of the published definitions.

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[11] Compare, for example, Allen (1936, pp. 18-47) and Krynine (1948, pp. 147-156).

The term *quartzite* is not used in this paper except as a part of already existing formational titles. It has been used in many previous reports on the southern Appalachians, but so loosely as to lack meaning, and without particular reference to the contents of quartz in the rock. The writer prefers the term *sandstone*, to which such modifying terms as quartzose, arkosic, indurated, and meta- can be added. If used, quartzite should be restricted to well-indurated quartzose sandstones such as those in the upper part of the Chilhowee group. Similar rocks have been called orthoquartzite by Krynine (1948, p. 149), but Allen (1936, pp. 38, 43) terms them silicinate quartz sandstones, and restricts the term quartzite to metamorphic rocks.

The terms *arkose*, *arkosic sandstone*, and *arkosic conglomerate* are used in the sense of Allen (1936, p. 44) for a sandstone containing 25 or more percent of feldspar, usually derived from the disintegration of acid igneous rocks. Those in the Chilhowee group contain more quartz than feldspar and very little of the other constituents of granite, such as mica and ferromagnesian minerals. They are well sorted and are probably subaqueous deposits. The term arkose has been restricted by Krynine (1948, p. 154), probably unjustifiably, to poorly sorted deposits, largely of continental origin, which occur in thick, lenticular sedimentary bodies.

The terms *graywacke*, *graywacke sandstone*, and *graywacke conglomerate* are used in the sense of Pettijohn (1943, pp. 941-943) for a primary sedimentary rock—a sandstone or conglomerate made up of grains of feldspar and quartz, with a large proportion of argillaceous matrix, and of characteristic gray or dark-gray color. Associated structural features such as intraformational conglomerates, graded bedding, and the great thickness of the deposit serve to set this rock type apart as a distinctive facies.

The term graywacke has been diversely defined, as indicated by Allen (1936, pp. 28-31). The Great Smoky conglomerate, which is here considered to be a graywacke in the sense of Pettijohn, has in the past been termed a graywacke on the basis of different definitions. Emmons and Laney (1926, pp. 15-16) refer to the primary sedimentary rocks of the Great Smoky as arkose and state that "intense regional metamorphism . . . produced from the arkose a graywacke." The term is therefore used by them for a metamorphic rather than a primary sedimentary rock. Moneymaker (1938, pp. 293-294) criticises this use of the term graywacke and points out that such rocks are more properly termed gneisses. However, he fails to recognize that the less altered phases of the Great Smoky are of graywacke facies, because he accepts Twenhofel's little used definition of graywacke as "the basic equivalent of an arkose."

Stose and Stose (1946A; 1947, pp. 632-634; also paper of 1949) have used a rock classification which, so far as the writer is aware, differs from all others previously discussed. Two of their formations of the Ocoee series are termed the "Hurricane graywacke" and "Great Smoky quartzite." In the Great Smoky Mountains, the term "Hurricane graywacke" is used by them for rocks here termed the 'Pigeon siltstone." Both in the Great Smoky Mountains and at Hurricane village, North Carolina, these rocks are fine-grained, laminated, indurated siltstones, in part greenish and chloritic, in part with dark carbonaceous seams. In the Great Smoky Mountains, the term "Great Smoky quartzite" is applied by them to rocks here termed the Great Smoky conglomerate, which the writer considers to be of graywacke facies. The use of the term "quartzite" for these quartz-poor, highly feldspathic rocks, would seem to be difficult to justify.

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ANDEAN UPLIFT AND EROSION SURFACES NEAR UNCIA, BOLIVIA

EUGENE H. WALKER

ABSTRACT Such knowledge of the physiographic development of the central Bolivian Andes as exists has been gained from a few rapid reconnaissances. Detailed studies in the region around Llallagua reveal a set of four erosion surfaces that show the stages of intermittent uplift of the Cordillera Physiographic and paleontologic evidence lead to the conclusion that the area was relatively low-lying toward the end of the Pliocene and that most of the uplift to the present elevations occurred during the Pleistocene There was slight mountain glaciation during the Pleistocene, but two terraces and stream bed alluvium suggest two stages of glacial climate. The region is now experiencing an increase in aridity.

INTRODUCTION

ALONG a section drawn east-west through central Bolivia (fig. 1) the Andes attain their greatest width, about 435 miles. This broad belt is composed of an eastern and a western cordillera separated by the Altiplano Basin. The rudiments of the geology of the Eastern Cordillera are known, for mining has been carried on at many widely separated points since the discovery of the Potosi silver deposit in 1545. Relatively little is known of the physiographic history. Moon (1939) gathered the scattered observations of many travellers and students but their data varies greatly in age and reliability. Gregory (1913) and Ahlfeld (1945) have discussed the origin of the spectacular gorge cut directly across the backbone of the Eastern Cordillera by the Rio La Paz. Bowman (1909, 1914) first noted the high level surface that truncates the Andean structures, and described the high shorelines that indicate the past higher levels of the Altiplano lakes, Poopo and Titicaca. Heald and Mather (1922), and Mather (1922) briefly mentioned a high erosion surface in the eastern part of the Eastern Cordillera. More recently Newell (1945, 1946) in the course of stratigraphic and structural work in the Lake Titicaca region, has clarified some of the events of the Tertiary.

With the intention of tracing the steps in the development of the Andean landforms the writer carried on field studies intermittently during the years 1942-45 in the region around Llallagua, where he was employed by the Patino tin mining interests. It is a pleasure to acknowledge here the aid given by

Marcelino Medrano of Cochabamba, Bolivia, who acted as muleteer, interpreter and guide on many a trip in canyon and high puna. In another sphere, the writer acknowledges his

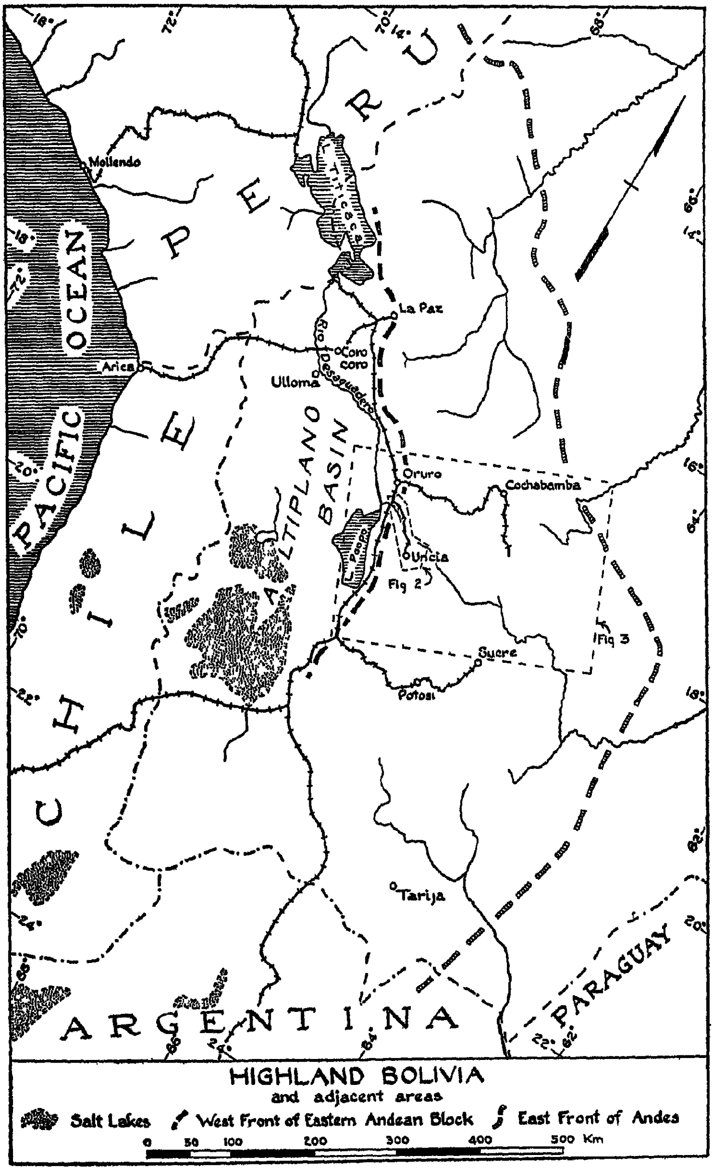


Fig. 1. Index map of Highland Bolivia.

indebtedness to Prof. Kirk Bryan of Harvard University for helpful criticisms and suggestions during the writing of this study. He also owes thanks to Prof. A. J. Eardley of the University of Michigan for his critical review of the manuscript.

GEOGRAPHY

The area studied lies a few miles east of the crest of the Eastern Cordillera, near Uncia (fig. 1), capital of the province of Bustillo. Llallagua and Catavi, the mine and mill settlements of the Patino Corporation, are nearby, and these towns are served by a railroad branching from the main line at Machacamarca (fig. 2). Machacamarca lies on a level plain which is an old lake bed, and a mile to the east the front of the Eastern Cordillera rises abruptly. The railroad climbs up the valley of the Rio Huanuni to the Paso del Bombo at 14,435 feet, where an extensive view is obtained of broad uplands of moderate relief, crowned by a few monadnock peaks. The total relief is somewhat more than 3000 feet, mainly due to deep valleys below the upland surface.

The elevation, coolness and semi-aridity of the area provide the steppe type of climate and the Andean bunch grass, *Stipa ichu*, is the strong dominant (pl. 1, fig. 2), though a carpet of shorter grasses usually is present between the clumps of ichu. At higher elevations a sparse growth of curious ground-hugging flowering plants gradually replaces the ichu, and in deep sheltered valleys some tall succulent grasses and a few shrubs appear.

Escobar (1943) shows the mean annual temperature to be about 9°C. (48°F.) and the precipitation about 400 mm. (15.7 inches). The precipitation is mainly as rain, from November to March. The mild winter then comes on and a couple of snowfalls of perhaps half an inch apiece may be expected, but hail is more frequent. Winter nights are frosty. Except during the rainy season the days are brilliant and cloudless.

The sparse vegetation permits rapid runoff so that torrents descend the minor watercourses during rainstorms, and all during the rainy season threads of water run in the braided channels that traverse the flat cobble-choked bottoms of the main valleys. In the succeeding dry season all minor streams

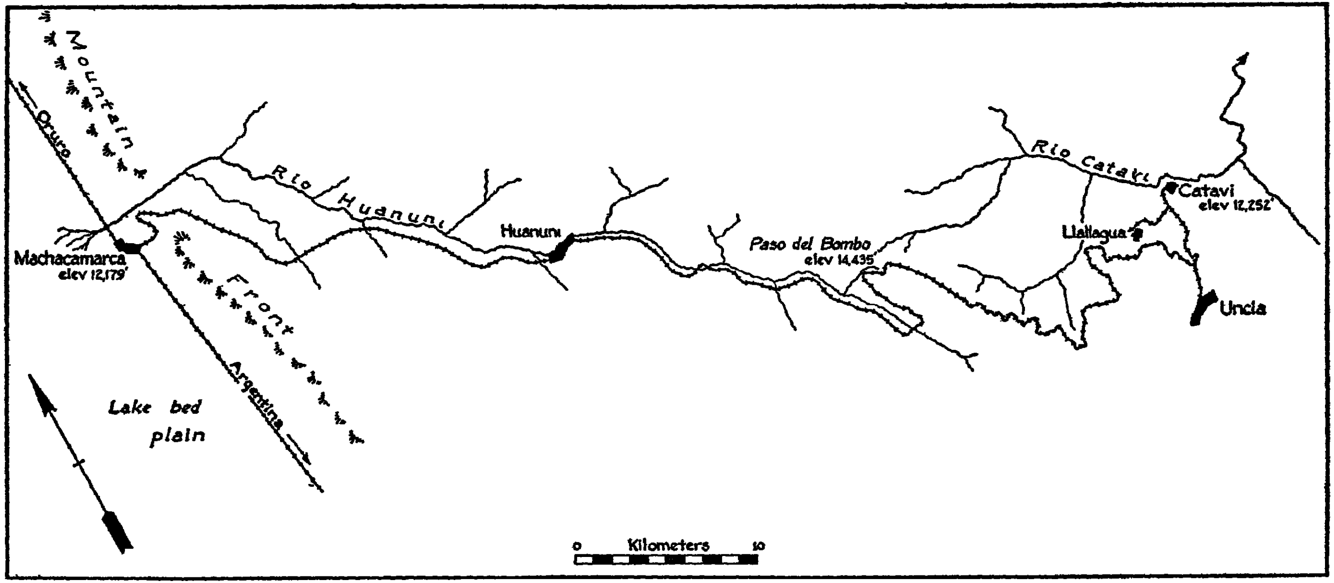


Fig. 2 Location map, Machacamarca-Uncia area.

are dry and the water gradually sinks out of sight beneath the surface of the alluvium in the main valleys.

The region is thinly populated by Aymara and Quechua Indians who have supported themselves since the most remote past by a simple pastoral and farming economy. The mining activity has introduced, more or less in the past 50 years, a town-dwelling population of about 30,000, mainly cholos (mestizos) of dominant Indian blood.

BEDROCK GEOLOGY

A section of about 6,000 feet of marine clastics and shales are exposed locally; they were first described by Koeberlin (1919). The strong Llallagua sandstone, about 200 feet thick, is a prominent ridge former. The rest of the sequence consists of greywacke, interbedded thin sandstones, and shales, all of intermediate to weak resistance to erosion. A few ill-preserved brachiopods found in a sandstone midway in the section are of Devonian age. Small patches of red continental sandstone and conglomerates lie with slight erosional unconformity upon the marine strata.

The Andean orogeny here creáted broad folds which strike northwest. Normally the folds are open, but stresses accumulated locally to cause overturned axial planes and thrust faults, both dipping steeply westward. A set of minor faults run northeast and cause short displacements of the northwesterly striking structures. A large normal fault runs north 30° west through Catavi and dips steeply west; the west side is dropped about 6,000 feet.

A volcanic neck, the Salvadora "stock," occurs along the axis of a large anticline. The cordierite-rhyolite of this body was greatly altered by the solutions that concentrated an enormous wealth of cassiterite locally.

Sheets of light gray tuff mantle broad areas. Though of similar composition to the rock of the volcanic neck, they lie on an erosion surface younger than the neck. The surface upon which they lie is undulating so that their thickness varies from about 150 feet to a feather edge.

The linear outcrops of resistant and non-resistant strata exposed along the fold axes have exerted the major physiographic control. The tuffs are resistant to erosion because of their light and porous nature which permits water to sink in rather than to run off. Thus the volcanics support mesas, usually bordered by cliffs (pl. 2, fig. 2) which develop because of columnar jointing in the volcanics and undermining the seeps.

PUNA STAGE

In the vicinity of the Paso del Bombo, the divide between the drainage eastward to the Amazon and that westward to the Altiplano Basin, there are patches of a terrane characterized by broad rounded shoulders with intervening swales and meadow-like expanses. This topography is at elevations from 14,200 to 14,500 feet and is localized upon the rocks of moderate to weak strength, between or near outcrops of the strong Llallagua sandstone which supports long ridges that rise 150 to 350 feet higher (pl. 1, fig. 1). Unfaulted blocks of sandstone cause a few monadnocks to rise still higher; the prominent peak Espiritu Santo has an elevation of 15,364 feet.

This surface is considered the equivalent of the Puna surface that McLaughlin (1924) described in the region near Cerro de Pasco, Peru, where

> "In general, the surface truncated the complex structures of the Cordillera with indifference, except here and there, where the course of a particularly resistant stratum is marked by a hog's-back ridge, or where a mass of intrusive rock forms an isolated hill or prominent group of hills of Monadnock type."

"Puna" is the Quechuan word for the high and cool grassy uplands throughout the Andes. The presence of this surface in Peru has been verified by all later observers, most recently by Harrison (1943), and Newell (1946).

In the Llallagua region the remnants of the Puna surface decrease in area and practically vanish with increasing distance from the divide. The long hogback ridges on sandstone which stood above the surface persist still and give the even skyline of mid and far distance.

Heald and Mather (1922) recognized remnants of what is probably the Puna surface in the highland southeast of Cochabamba, and wrote that

> "During the late Tertiary (?) this area was a low-featured peneplain, which is now represented only by ridgetops and a few undissected remnants."

Still further south, in the sierras immediately west of Santa Cruz, Mather (1922) saw

> " . . . occasional upland surfaces with gentle slopes which

> bevel the upturned sediments and are ordinarily surmounted by slightly higher rounded knobs."

The high mature surface that Bowman (1909) noted on the border between Bolivia and Chile at the crest of the Western Cordillera inland from Iquique, may also be the Puna surface.

Fragmentary as the evidence is, it is apparent that throughout a vast area of the Bolivian and Peruvian Andes a distinct surface bevels structures and intrusives. The surface is probably broadly arched at present. The traces of it are lost in the angular scenery of the Eastern Cordillera before it has declined much in elevation, and it is possible that its remnants end against a line of great faults that Bowman (1909) and Mather (1922) consider to separate the mountainous mass from the Amazon lowland. McLaughlin (1924) states that in western Peru the surface is ". . . obliterated by erosion before it descends much lower than 12,000 feet . . . ." The surface that Bowman (1909) described in the Western Cordillera slopes gently toward the Pacific beneath a thickening cover of late Tertiary sediments and volcanic effusives.

Figure 3 presents the rectangular drainage pattern on the eastern side of the Eastern Cordillera. Lengths of subsequent streams have taken northwesterly courses parallel to the folded structure but long stretches of the major streams run northeast across the regional grain. These transverse links presumably are remnants of the drainage of the Puna stage. It is usual to consider such transverse drainage to have developed by superimposition from a cover such as the coastal plain sediments furnished on the east slope of the Appalachians. Since no traces of such covering sediments exist upon the Puna surface, the relations are most readily explained by assuming superimposition from a cover onto an old age surface which existed above the Puna surface, but which was here wholly destroyed in the making of the Puna surface. Such a surface may eventually be found beneath covering sediments in the subandean foothills to the east of the zone of uplift. The subsequent streams, in any case, testify to a long history of differential erosion and piracy since Puna time.

VALLE STAGE

Some 900 to 1,000 feet below the remnants of the Puna surface is a terrane developed broadly across the upturned sediments. Along the main drainage divide massive ridges exist above this level but elsewhere only occasional monadnock ridges survive along parts of the outcrop of the strong Llallagua sandstone. The landscape at this level is one of valley floors, usually very wide, above which mountains rise 1,000 to 1,400 feet.

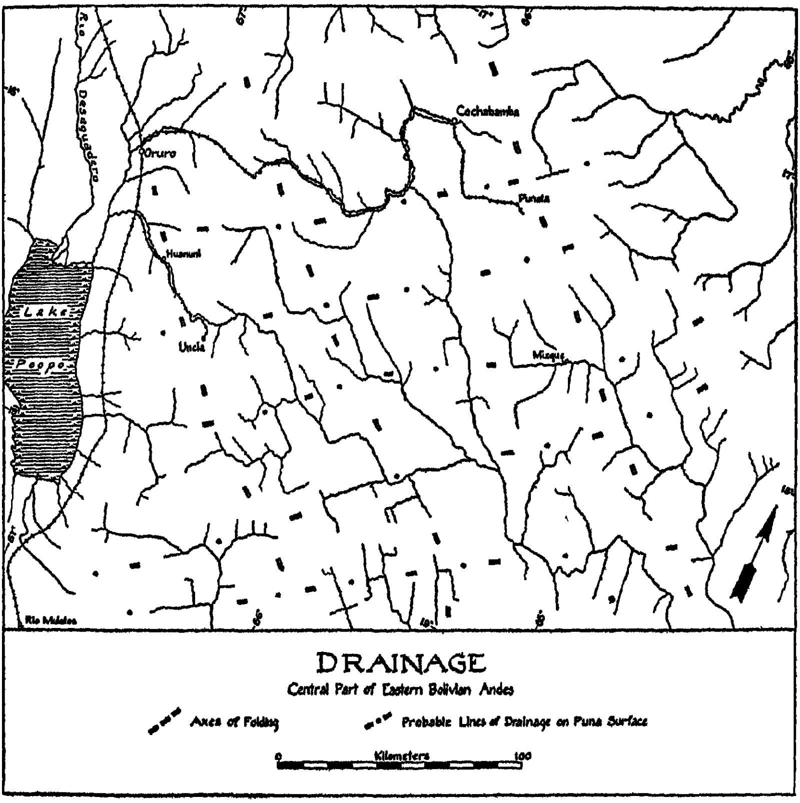


Fig. 3. Drainage of central part of Eastern Andes.

Volcanics, mainly firm tuffs, mantle much of this surface. In old valleys they may be 150 feet thick and they thin out upward on the old slopes. Usually the volcanics lie on bedrock that is weathered to a depth of several feet but in some places one may find stream gravels of quartzite marking buried drainage courses. The volcanics have diminished the local relief to produce the mildly rolling terrane typical of this main upland horizon (pl. 1, fig. 2).

These features record the following sequence of events. The slow course of erosion at the Puna level was interrupted by uplift (and further west probably by faulting, as will be argued) which caused the streams to lower their courses approximately 1,000 feet before coming to grade. Long stillstand allowed reduction of most of the area to the new level. Weathering of bedrock and the existence of only resistant rock types, mainly quartzite, in stream channels testify to ample chemical weathering at this time.

Probably this interval of valley broadening should be correlated with the Junin stage that McLaughlin (1924) described, for in Peru, after the long stillstand of the Puna stage "An early, rather slight uplift, followed by a long period of quiescence, allowed the development of a series of broad valleys with flat gradients and gentle side slopes." The type locality, the Junin Valley, provided a name for this surface and stage. Since 800 miles separate the localities studied in Peru and Bolivia the correlation is uncertain and the present writer applies the name Valle to the surface in Bolivia. This Spanish word refers to broader and more mature drainage basins than the English word valley, and seems appropriate for the open terrane that resulted from the erosion of this stage.

CHACRA STAGE

About 500 feet below the Valle surface, or slightly less than halfway down to the present stream beds, there are traces of a surface. Along the valleys of main streams the vigor of later erosion has left of it only a few flat-topped spurs and occasional gravel-strewn benches. Larger tracts, still little dissected, persist along the courses of those minor streams and headwaters that had the fortune to lie upon the less resistant bedrocks. Such an area is illustrated in plate 2, fig. 1. Another prominent trace of this surface (pl. 2, fig. 2) owes its preservation to a piracy which greatly weakened the stream that crosses it; the stream is now intermittent and so incompetent that even though it crosses soft shales a hanging junction has developed where it joins the main stream.

This surface is important chiefly as evidence of a pause in the major downcutting of post-Valle time. The volcanics that blanket the Valle surface were perhaps the surface manifestation of deep seated disturbances accompanying a regional uplift that brought about renewed downcutting. In some cases the slopes beneath volcanic cover incline gently toward the present valleys, indicating that the main streams were able to hold to their courses, though the pattern of the minor streams was effaced. Downcutting continued for about 500 feet, at which point a pause allowed valleys to be broadened upon softer bedrock.

This surface seems likely to correlate with the benches McLaughlin (1924) described in Peru midway between the uplands and the valley floors and is given the same name. "Chacra" signifies a small plot of tilled ground; here in Bolivia as well as in Peru much of the cropland of the Indians is at this level, below the bleak uplands and above the generally stony valley floors.

PIEDMONT STAGE

A very obvious surface of planation lies 400 to 500 feet below the Chacra remnants and 150 to 200 feet above the stream beds. Along those portions of the valleys floored with shales and thin-bedded sandstones are expanses as much as several miles wide, the "pampas" of the region. These old valley floors constrict to canyons at the outcrops of the Llallagua sandstone. No counterpart of this surface has been described elsewhere and it will be referred to as the Piedmont surface since it forms so prominent a footslope below most of the local relief.

It is characteristic of this surface that bevelled bedrock inclines toward the valley centers (pl. 3, fig. 1). Where bedrock is mantled with alluvium (pl. 3, fig. 2) the sloping bedrock can be seen in the walls of the quebradas or gulches that trench below the alluvial cover. We have here a rock pediment rather than a surface of lateral planation. The alluvium reaches a thickness of 100 feet in some places but always thins to a frayed edge toward the hill and mountain slopes.

At its base the alluvium is coarse and cobbly, but not badly sorted; low grade placers of cassiterite have encouraged some prospecting. Upward from the base the grain size at first decreases slightly; then a transition sets in and particle size increases till the alluvium becomes even bouldery. Some of the stones are ill-rounded, many of them bear the percussion marks showing handling by torrential streams. This coarse material grades unevenly upward through finer grained zones and at the top of the section there are lenses of sand and silt.

At one locality there is in the upper part of the sequence a thin bed of peat. It is reported that a number of years ago the remains of a mastodon were found in the higher part of the alluvium.

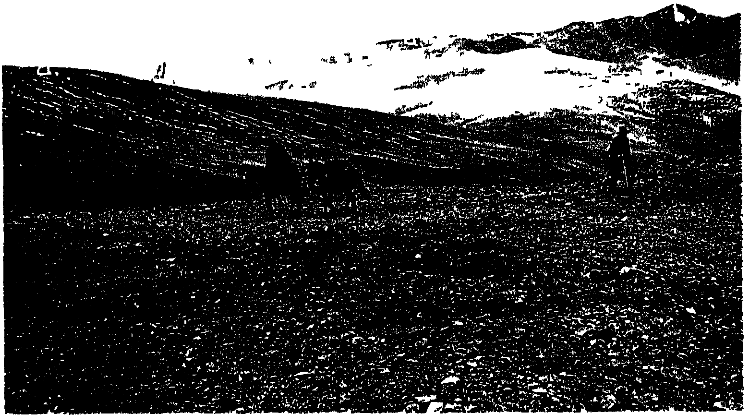
The significant downcutting from the Chacra surface indicates resumption of the spasmodic Andean uplift, and the pediment surface strongly suggests drier conditions than previously. The lower alluvium can probably be considered a normal accumulation on a pediment graded to an incompetent stream, but some explanation must be found for the change midway in the alluvial section.

The coarsening of the alluvium apparently records accelerated erosion of slopes and delivery of excessive loads to lowland streams. The writer believes that this was due to the onset of glaciation in the Andean region. Snowline was high here, for only above 13,500 feet are there striated outcrops and there are no obvious moraines. Waste was not moved down to the valley floors by ice. However the colder climate undoubtedly increased the available supply of waste because of frost breaking of rocks and decreased vegetation cover, and solifluction would hasten material downslope. Valley trains of alluvium were spread.

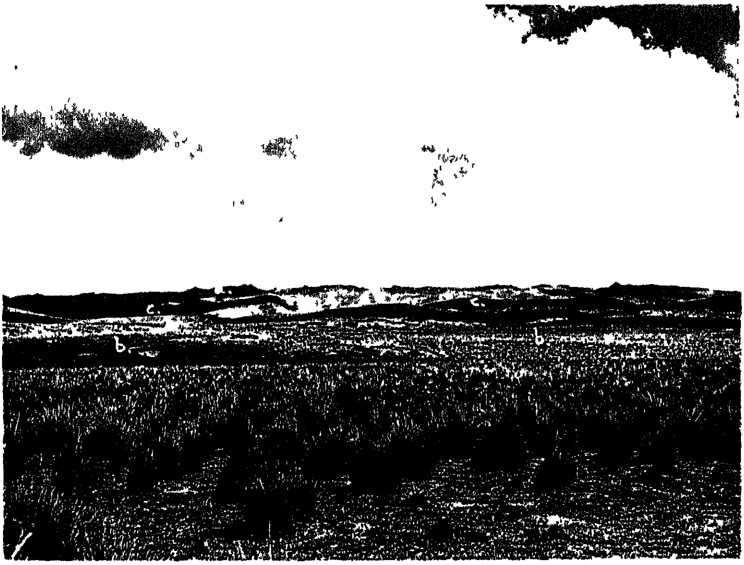
During the waning stages of glaciation finer sizes of sediments were deposited. The mastodon remains and the patchy peat deposits indicate a cool and damp climate toward the end of this period, with broad marshy valley flats traversed by braided or meandering streams.

LATER PLEISTOCENE STAGE

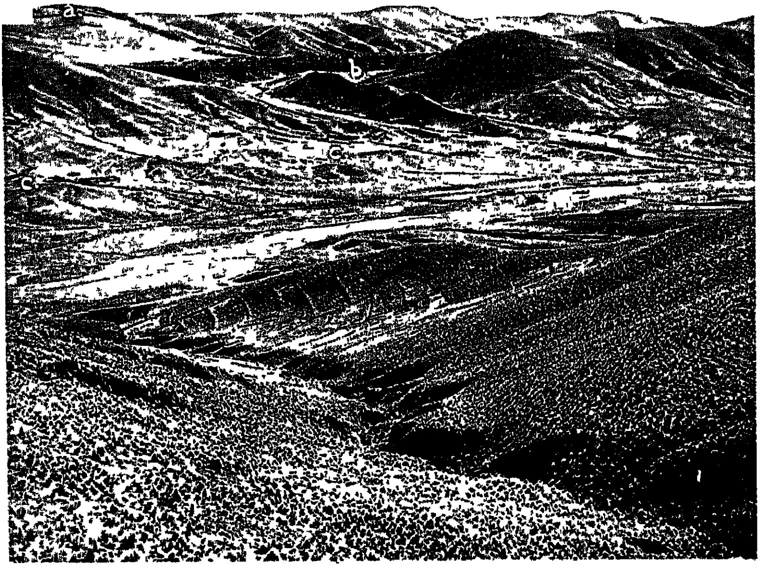
Below the Piedmont surface some 200 feet or less there are flat valley floors strewn with coarse alluvium (pl. 3, fig. 2). West of the divide and downstream from the Huanuni mine the valley of the Rio Huanuni is at places more than a mile wide and the gravels are so thick that a large floating dredge is employed to work them for their cassiterite content. This particular sheet of alluvium was clearly graded to the level of a lake which occupied the Altiplano Basin to the west and which left prominent shoreline features about 115 feet above the basin floor. This was Lake Minchin; lake bed sediments and the delta deposits lain by the Rio Huanuni have been exposed near Machacamarca by later stream cutting.



REMNANTS OF THE PUNA SURFACE
Foreground elevation about 14,400 feet Puna surface at (a), monadnock on Puna surface at (b)

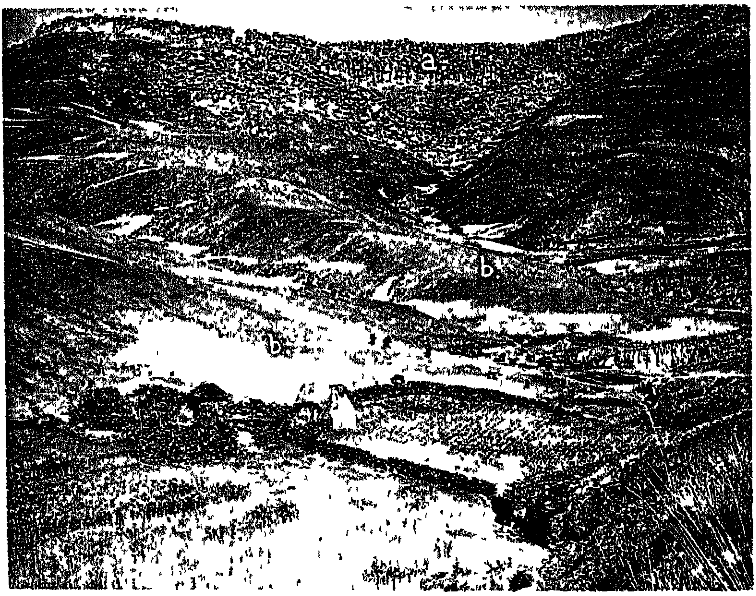


VALLE SURFACE MANTLED WITH VOLCANICS
Foreground elevation about 13,500 feet Puna ridgetop at (a), upland on volcanic mantle at (b), bedrock hills of Valle surface at (c)



HEAD OF HUANUNI VALLEY

Foreground elevation about 14,300 feet Level skyline at (a) is Puna surface, at (b) a portion of the Valle surface preserved beneath volcanic cap, the Chacra surface, little dissected is at (c)



CHACRA VALLEY

Volcanics overlying Valle surface at (a) The light colored zone marks the base of the volcanics The Chacra surface at (b) has here an elevation of about 12,700 feet

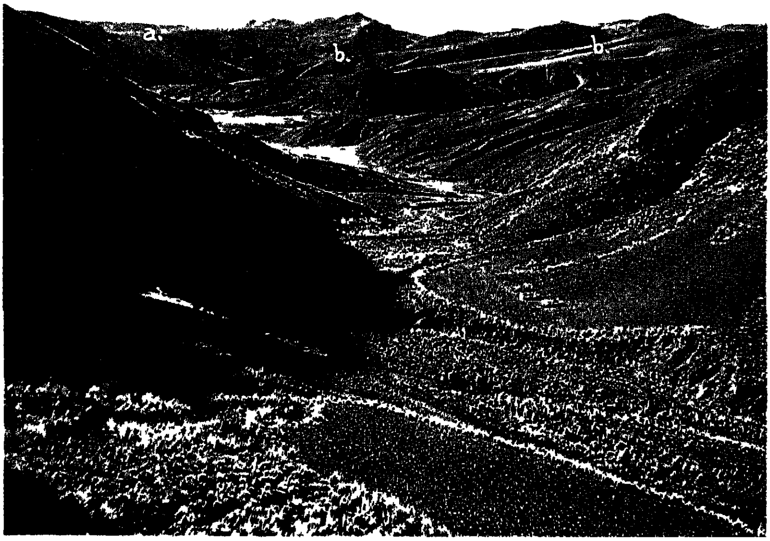


FIGURE 1 MIDDLE COURSE OF HUANUNI VALLEY
Elevation of immediate foreground is about 13,000 feet At (a) is a large remant of the Chacra surface, at (b) the Piedmont benches cut across folded shales and thin sandstones

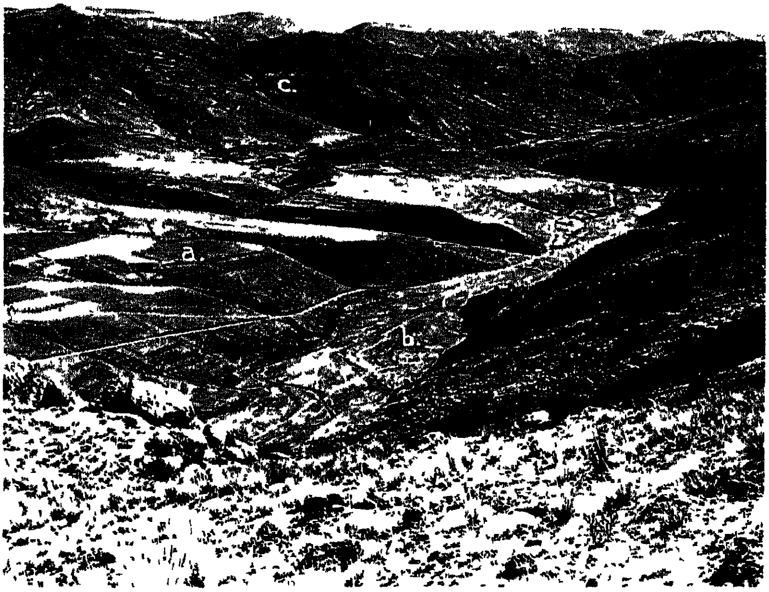


FIGURE 2. VALLEY OF RIO CATAVI
At (a), elevation 12,400 feet, the surface of the alluvial fill on the Piedmont surface, at (b), the braided channel of the Rio Catavi Immediately above (c) is a small patch of the Chacra surface, downstream from a gorge cut across the Llallagua sandstone

The trenching of streams below the Piedmont surface began after the first glaciation, and locally the streams lowered their courses as much as 100 feet into the bedrock beneath the alluvium. Renewed alluviation seems to have been due to a second period of glacial climate. The deep fill in the lower part of the Huanuni valley evidently is due to the fact that the gradual rise of Lake Minchin lifted the base level of the Rio Huanuni at least 115 feet and caused aggradation far upstream. East of the divide streams are locally graded to strong beds outcropping in their channels and ultimately to the Atlantic, so local aggradation was due to overloading during glacial climate.

RECENT EROSION

Due to the disappearance of Lake Minchin, the Rio Huanuni and other streams draining westward to the Altiplano Basin have begun to trench their valley fill. The nickpoint in the Huanuni valley has advanced only about three miles upstream. In general, erosion in main stream beds is taking place very sluggishly due to coarseness of the load to be removed, and small supplies of water. East of the divide, however, the great drainage systems slowly concentrate many trickles of water into single channels, with the result that erosion increases rapidly downstream.

At the present, hill slopes are being gullied, and small arroyos are being cut in the alluvial floors of many tributary streams. Here most of the steep slopes are very thinly covered with bunch grass, others are almost totally barren. Without doubt this erosion is partly due to the land use tactics of the Indians. The search for fuel is intense; hardly a bush survives after its stem attains the thickness of a lead pencil. The bunch grass may be hauled up roots and all for fuel or thatching. The more palatable grasses are cropped closely by sheep and llamas. The marvellous and laboriously built andenes or rock walled terraces built in ancient time still persist on the hill slopes, but the narrow strips of soil they provide are now neglected in favor of larger fields of considerable slope. The accelerated erosion is therefore partly induced by land use.

It is proper, however, to point out the evidence of increasing aridity. The level of Lake Titicaca in the northern part of the Altiplano has been falling slowly for decades, so that the approaches to steamship wharves have had to be dredged repeatedly. The volume and depth of the Rio Desaguadero, the outlet of Lake Titicaca, have notably decreased with the result that Lake Poopo receives less water and has shrunk greatly. Slope erosion in Bolivia, as in the southwestern United States, is due probably to a combination of climatic change and severe use.

SUMMARY OF HISTORY AND DATING OF EVENTS

In the area under discussion paleontologic evidence for dating the surfaces and events is absent so it is necessary to establish correlation with areas where such evidence, scanty as it may be, does exist.

The cutting of the Puna surface is the first event for which a date is required. The surface truncates the folds and intrusives of the Andean region so the date of the orogeny sets a maximum age limit to the surface. Unfortunately the date of the orogeny is still disputed. Berry (1919) assigned the orogeny to the late Pliocene because he dated as Pliocene the plant leaves that are found in the folded sediments at Coro-Coro in the northern Altiplano of Bolivia. At Potosi, sediments with similar leaves are little deformed, but are cut by a volcanic neck.

Near Ulloma, in the broad valley of the Rio Desaguadero south of Lake Titicaca (fig. 1) the slightly consolidated "Puna" beds lie essentially horizontally upon a surface that bevels the older folded sediments. The surface is probably the Puna and in any case it can hardly be older. The Puna beds contain a rich mammalian fauna of *Hippidion, Mastodon, Scelidotherium, Mylodon,* and *Megatherium,* which Philippi (1893) dates as Upper Pliocene. This occurrence suggests that the Puna surface was fully developed at or before the end of the Pliocene.

The evidence from these two localities is conflicting, or at best allows only a fraction of Pliocene time for the folding, intrusions, uplift and long erosion. The recent tendency has been to ignore the date assigned to the fossil leaves and to make the orogeny earlier than Pliocene. Turneaure and Welker (1947) believe that the intrusives date from some part of the interval between the Eocene and the Pliocene;

Steinmann (1929) has shown that much if not all the folding in Peru is early Tertiary, and Bruggen (1934) finds a similar date in Chile. Ahlfeld (1941) writes

> "Groeber maintains that the principal Andean folding in northern Argentina took place in the interval between the Oligocene and the Miocene. As long as the exact determination of the age of the folding in Bolivia is not possible, we are going to accept that this took place at the same time as in northern Argentina."

In his recent paper Newell (1946) refers the latest Andean folding to the Miocene.

The evidence is therefore fairly strong that the erosion of the Puna stage took place during a large part of Pliocene time and the surface was completed and locally mantled with the Puna beds before the end of the Pliocene.

The undisturbed Puna beds raise a problem as to their origin and relations to the later erosion surfaces. Patches of these sediments occur throughout the Altiplano Basin. At Ulloma, Ogilvie (1922) reported " . . . reddish and yellowish sands with irregular lenses of gravel and occasional marl and clay bands . . . ," while nearby at Mauri Ahlfeld (1941) mentioned rose colored marls and sands and conglomerates. In the valley of La Paz the writer has seen loose red and white sands interbedded with conglomerates containing pebbles from the Tertiary granites not far to the east.

Similar beds lie in local basins within the Eastern Cordillera, at Cochabamba and at Tarija, according to Oppenheim (1943). In the southernmost part of Peru, just north of Lake Titicaca, Gregory (1914) and Eaton (1914) found and described basin sediments of the same type. Their locality is near Ayusbamba in the Upper Apurimac Basin, where loosely consolidated sands, clays and marls lie with great angular unconformity on folded bedrock. The beds are remarkable for their fauna of *Camelidae*, *Equidae*, *Cervidae*, *Elephantidae*, and *Mylontidae* of late Pliocene or early Pleistocene age.

These occurrences have in common their considerable altitude, ranging from 8,500 feet at Cochabamba to 12,300 feet at Ayusbamba, the great unconformity separating them from the older rocks, and their lithology. They are the clastics deposited in stream channels, on flood plains and deltas, and the marls and clays of lakes and swamps. Red beds are common. Drainage was to local basins rather than through-going. The climate was warm and the land low-lying if we take the red beds to signify tropical or sub-tropical weathering of the source areas, and clearly the herbivores required abundant vegetation.

The change from the through-going drainage upon the Puna surface to the basin drainage indicated by the Puna beds is most reasonably explained by assuming that block faulting took place. Moon (1939) first suggested faulting as an explanation for the basins, although McLaughlin (1924) had earlier considered that " . . . some warping and possibly block faulting . . . " ended the Puna stage in Peru. Newell (1946) agrees that the Altiplano Basin of Bolivia is probably a broad crustal segment downdropped with respect to the Eastern Cordillera. The relatively straight west front of this Cordillera is the much eroded fault line scarp.

The new relief provided by this faulting set erosion to work upon a weathered terrane. Part of the detritus accumulated in the Altiplano Basin and the smaller intermontane basins to form the Puna beds, part was removed eastward toward the Amazon Basin. The east-running streams gradually opened out the Valle surface, to be dated as Plio-Pleistocene because of contemporaneity with the Puna beds.

As previously mentioned, lithology and fauna of Puna beds indicate low-lying land. The subsequent uplift, which amounts to about ten or twelve thousand feet in some places, must have taken place in the Pleistocene. A similar conclusion as to the recency of the Andean uplift was obtained by Harrison (1943) from his work in Peru. If one million years are allotted to the Pleistocene (Zeuner, 1945), the rate of uplift if steady would have been about 0.01 feet per year. The post-Valle erosion surfaces show that uplift was not continuous, so the rate must have exceeded this figure considerably at times.

The volcanics emitted in the Llallagua region at the close of the Valle stage apparently have their counterparts throughout the Andean region, for vast areas around and east of Potosi are thickly blanketed. To the north volcanics are still widespread, but much thinner. At Mauri the Puna beds are overlain by tuffs, and dacitic ash is mingled with the upper beds at Ayusbamba. It is probable that the great volcanic piles of the Western Cordillera began to form at this time. The region-wide vulcanism, as well as the climatic change due to uplift, was no doubt a factor contributing to the extinction of the herbivores of the Puna beds.

The Chacra surface is early Pleistocene and significant only as evidence of pause during uplift. The rock-cut floor of the Piedmont surface is a pediment and gives the first evidence of semi-arid climate. This climate may have been due merely to elevation and position far inland; it may also mean that by this time the Western Cordillera had grown high enough to further modify the inland climate.

The nature of the alluvial fill upon the Piedmont surface furnishes evidence of the first of the two glacial stages recorded in the Andes. This stage should be recorded by high shorelines in the Altiplano Basin, but in the southern part of the Basin no such shoreline except that of Lake Minchin is evident. Up to 100 meters or 330 feet above Lake Titicaca are the shorelines of Lake Ballivian described by Bowman (1914) and others (see Moon 1939). Newell (1946) believes that the first lake to occupy the Titicaca basin was pre-glacial, but some of the beaches above the present lake may eventually be shown to illustrate the first and second glacial stages.

In the Llallagua area an interglacial is clearly evident in the interval of erosion that separates the two stages of alluviation, and the second stage of alluviation has, in the case of the streams draining to the Altiplano, been directly correlated with the Lake Minchin beach.

The climate is at present suitable for the formation of pediments and this is the form now developing west of the divide in response to the more or less stationary base level furnished by the Altiplano Basin. East of the divide the slow pace of stream-bed erosion is due to be quickened fairly soon. The Rio Catavi, which may be taken as typical of east-running streams, has an overall gradient of 131 feet per mile in the first 35 miles of its course downstream from Catavi. Canyons will be cut before grade is reached. The divide between east- and west-running drainage is destined to shift westward until ultimately the Altiplano Basin will be drained eastward to the Amazon. This stage of drainage development has already been reached in the northern part of the Altiplano Basin;

the gorge in which La Paz lies is a trench excavated in the soft fill of the Basin by a headwater of the Amazon.

CONCLUSION

The principal aim of the writer has been to establish a satisfactory outline of the events of the late Tertiary in the Eastern Cordillera of Bolivia. This has involved reference to other localities about which a meager amount is now known. A framework of reference has been provided, however, for the scattered observations that will slowly accumulate, and perhaps for further systematic studies in small areas. The high Andes offer a practically virgin field for geomorphic study; the stages of glaciation and the glacial lakes are perhaps the most important problems inviting study.

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THE JOURNAL CHANGES EDITORS

With publication of the July number this year, the name of Richard S. Lull, Editor for many years, disappeared from the cover of this Journal. Logically that issue should have carried an explanation of this event, about which most of our readers had no advance notice. However, Professor Lull's successor was just returning to New Haven and adjusting his schedule after nearly a year of field work in the West. Moreover, in midsummer large numbers of the Journal family are away from their reading tables. September, beginning the season of return to indoor tasks, is an appropriate month for this announcement.

Professor Lull became Editor-in-chief of the Journal at the start of the year 1933, just three and a half years before retirement from his professorship at Yale. Thus he kept the editorial post for sixteen and a half years. Up to 1938 he exercised full duties as Director of the Yale Peabody Museum; and he still occupies his office in the museum building, carrying on his research in vertebrate paleontology. He relinquishes the editorship, but retains his keen interest in science and in the Journal.

In looking through a full set of the American Journal of Science, which has run without interruption from 1818 to the present, we discover only three predecessors of Professor Lull whose names have appeared in first place on the front cover. These men, with dates of their incumbency, are as follows: Benjamin Silliman, Sr. (founder of the Journal), 1818-1870; James Dwight Dana, 1871-1894; and Edward S. Dana, 1895-1932. In fact, however, Alan M. Bateman and Ernest Howe performed full duties as Editor, the former in the period 1925-26, the latter from 1927 through 1932, although both these men chose, as a courtesy to Professor E. S. Dana, to print his name ahead of their own. Furthermore from 1838 through 1864, Benjamin Silliman, Jr., shared the editorial duties, and his name followed his father's on the cover. In the period 1846-1864 the name of J. D. Dana was added to form a triumvirate that endured until the younger Silliman withdrew. The elder Dana continued to assist Benjamin Silliman, Sr., through 1870; and during much of the period 1871-1894, the name of E. S. Dana was carried with that of his father. Actually, therefore, six men had served as editor of the JOURNAL during the 115 years ending in December, 1932.

Professor Lull assumed the editorship at a difficult time, in the worst part of the economic depression. His administration brought the JOURNAL through that period, and also through the hard years of the war when subscription lists suffered and production of scientific papers was at low ebb. He comes of a family with a long naval tradition, and in his jovial parlance he has now "turned over the ship." In the same spirit we salute him as a gallant skipper who has brought the ship through fair weather and foul, and has turned it over in seaworthy condition to a new command.

John Rodgers, who began as Assistant Editor in the fall of 1948, continues in that capacity, thus supplying an effective link between the two administrations. To readers and contributors, the new editors promise their best efforts to maintain the high standard of service set by Professor Lull and his predecessors.

CHESTER R. LONGWELL.

REVIEWS

The Structure of Matter; by FRANCIS OWEN RICE and EDWARD TELLER. Pp. xiii, 361. New York, 1949 (John Wiley and Sons, Inc., $5.00). In this work, the authors attempt "to acquaint the reader with the scope of the phenomena that can be explained with the help of quantum mechanics." Their aim has been to accomplish this with the minimum use of mathematics, and even the Schroedinger equation was omitted on this account. The result is a book which should be of real value to those who wish to gain a qualitative feeling for the quantum mechanical approach. For many applications, this may be all that is necessary, since the problems are so complex that the difficulties of a rigorous, mathematical treatment are insuperable.

The book is outstanding for the way the authors have succeeded in presenting concrete models for the often tenuous abstractions of the formal theory. They have accomplished this by a liberal use of diagrams the significance of which is carefully explained in the text.

The range of material covered is wide, ranging from the structure of the hydrogen atom to the sources of energy in the stars and the genesis of the chemical elements. Among the thirteen chapters are treatments of the periodic system, the chemical bond, intermolecular forces, the interaction of matter with electric and magnetic fields, molecular and atomic spectra, nuclear chemistry, and the solid state. As might be expected in an attempt to cover so wide a range of subjects, some of the topics are not covered in so refined a manner as might be desired by a specialist in the particular field. In many cases, however, quite sophisticated concepts are clearly presented. A notable example is the treatment of the modern theory of metals.

It is unfortunate that a book such as this, which is intended as an introduction to a large number of specialized fields, should include almost no references to works to which the reader can turn for more information about some specific phase of the subject. A selected list of general references for each chapter would increase the value of the book considerably. The book is also somewhat more limited in its scope than its title would indicate in that it treats only the properties of the atoms and molecules of which matter is composed. The way in which molecules cooperate to produce macroscopic effects, as discussed in the kinetic theory of matter and statistical mechanics, comes in only incidentally. This limitation excludes any coherent explanation of the properties of liquids and gases.

The book will be of interest to almost all scientists, and will be of particular value to those working in fields in which the rigorous application of quantum mechanics is impractical, but in which the quantum mechanical method of thought may furnish a useful guide in interpreting the observed phenomena. HENRY C. THACHER, JR.

Sedimentary Rocks; by F. J. PETTIJOHN. Pp. xviii, 526; 40 pls., 181 figs. New York, 1949 (Harper & Brothers, $7.50). Professor Pettijohn's contributions to the study of sedimentary rocks are many and important. One thinks at once of his studies of Pre-cambrian sediments in the Canadian shield and of the widely used *Manual of Sedimentary Petrography* prepared jointly with W. C. Krumbein. His new book on sedimentary rocks is therefore a welcome addition to the literature of the field.

The book contains a great deal of valuable information, grouped around three major themes: properties of sedimentary rocks (chapters 2-5), kinds of sedimentary rocks (Chapters 6-11), and processes that form sedimentary rocks (Chapters 12-15). The line drawings are many and clear, and the constant use of triangular and other quantitative diagrams continues the excellent trend established in the *Manual of Sedimentary Petrography.* The 40 collotype plates, showing specimens, outcrops, and especially thin-sections, are among the finest illustrations this reviewer has ever seen in a textbook. The book itself is a fine example of good book-making.

Unfortunately, the material presented in the book shows signs of hasty compilation instead of the thorough working and reworking, rearrangement and checking, that the reader has a right to expect of the author. Some of the results are merely vexing, such as inadequate proof-reading, as in references and technical terms, frequent lapses in grammar, and poor organization. More serious is the failure to make various parts of the book consistent and to digest the material presented so that a clear summary of existing knowledge (and ignorance) emerges. On several subjects, for example oolites in Chapter 2 and abrasion and sorting in Chapter 13, the student is overwhelmed with pages of (frequently contradictory) results from the literature, but the subject is never summarized or generalized by the author. Not that the author needed to generalize arbitrarily among the contradictory data; even a summary of the contradictions and their implications would have been very helpful. As it is, unexplained contradictions abound (there are four on page 269 alone), and though doubtless all of these can be explained without difficulty, it is unfair to ask the elementary student to figure out each one without help from the author.

The fine aspects of the book, however, such as the great number of references scattered as footnotes through the text, will make it a good source for more advanced students. The last two chapters (on deposition and lithification) are especially good and are freer from errors (though even here fig. 116 contradicts the adjacent text and fig. 115 its own caption). Chapter 14 (Deposition) is a competent statement of the current tectonic interpretation of sedimentary sequences, and here the author has not been afraid to generalize (though as a result one reads that carbonate rocks are virtually unknown in the "flysch" facies whereas they form a considerable part of the type Flysch, and that the Alpine Molasse is red whereas it is gray and green). There is so much good material here, and Professor Pettijohn is so well qualified to present it, that one can only hope that he will take the necessary time to revise the book with care, checking throughout for inaccuracies and contradictions, and digesting and summarizing the rich but chaotic literature, so that students will find the book a safe and adequate guide. JOHN RODGERS

Map of the Americas, 1:5,000 000 (5 sheets, each 46 x 35 inches), $20.00. *Map of the World*, Miller cylindrical projection, 1.30,000,000 (57 x 35 inches), $2 50. American Geographical Society, New York.—The American Geographical Society has now completed its 1:5,000,000 map of the Americas, of which the three Latin American sheets were published in 1942. The map, plotted on a conformal double conic projection especially computed to procure minimum distortion throughout the Americas, is a fine example of the Society's high cartographic standards. Layer tints on land and sea emphasize the physical relief, and the principal streams, towns, roads, railroads, and airfields, and the country and state or province boundaries are shown. All available information was canvassed, and in places, as in the Canadian Arctic, the map shows data not hitherto shown on any map of like scale. Indices to all names given on the map are available.

The Society has also released recently a map of the world (except for Antarctica), showing land elevation (by six layer tints), the principal rivers and towns, and international boundaries (mainly as of 1947). This map also is on a new projection, which avoids the extreme expansion in high latitudes inherent in the Mercator projection, but at the expense of very considerable distortion, A diagram on the reverse facilitates determination of great circle distances and routes. JOHN RODGERS

Klockmann's Lehrbuch der Mineralogie; by PAUL RAMDOHR. Thirteenth edition, Pp. xii, 674. Stuttgart, 1948 (Ferdinand Enke Verlag, DM 50.00).—This excellent textbook of mineralogy brings to date a long series of editions that have appeared under well known names. The book contains two major divisions, (I) general and (II) descriptive mineralogy. General mineralogy is subdivided into (1) physical mineralogy, including crystal form, crystal fine-structure, and physical properties; (2) chemical mineralogy and crystal chemistry; (3) formation of minerals in nature, including geologic setting, mineral deposits, paragenesis, and an interesting list of 81 famous mineral localities with the important and characteristic minerals of each; (4) mineral uses. The descriptive division is conventionally divided into sections based upon a crystallo-chemical classification of minerals, largely following Strunz's *Mineralogische Tabellen* (1941).

In briefly reviewing so inclusive a book, only a few points of special interest can be selected, other equally important ones being necessarily omitted. Ramdohr's crystallographic nomenclature could be made more useful by the inclusion of the Hermann-Mauguin class symbols and space-group symbols instead of (or in addition to) the Schoenflies symbols which adequately describe only the 32 classes (serial numbers for space groups isomorphous with each class can hardly be considered "adequate"). In general, the approach to crystallography and crystal structure is modern and satisfactory It is somewhat heavily weighted in favor of morphology.

It is not satisfactory to find more than one system of nomenclature for the principal refractive indices The symbols used in the general descriptions of methods and principles are the widely-known Greek letters α, β, γ, ϵ, ω. In some of the mineral descriptions we find the same symbols; but in other descriptions we note n_o, n_e; and in still others we find n_ω, n_ϵ, n_α, n_β, n_γ. There may be some justification for n_o and n_e—n clearly implying *refractive index,* and the subscripts indicating the name of the ray or vibration direction involved. But the use of the Greek letters as subscripts after defining them in accordance with common German usage as the refractive indices for certain rays, seems most undesirable. If n is to be used, its subscripts should refer to the vibration axes *(either x, y. z. or* $\mathfrak{a}$, $\mathfrak{b}$, $\mathfrak{c}$,) *should not be the symbols* that independently denote refractive indices. Least desirable of all is the apparently indiscriminate mixture of several systems of symbols. HORACE WINCHELL

Minerals and How to Study Them; by the late EDWARD S. DANA, revised by Cornelius S. Hurlburt, Jr. 3rd Ed. Pp. x, 323; 387 figs., colored frontispiece. New York, 1949 (John Wiley & Sons, $3.90).—This book, intended specifically for beginners in mineralogy who lack the scientific background to attempt the more comprehensive texts, will be welcomed as a timely revision of a book already popular in its field It is the least advanced in the Dana series of mineralogies, which includes also the *Manual,* the *Textbook,* and the *System.*

Doctor Hurlburt's revision has consisted chiefly of three things: first, rephrasing of the late Mr. Dana's words in a style better suited to the present day, while not losing a bit of their cogency and clarity; second, introduction of new material which has come as a result of over half a century of scientific progress; third, complete rewriting of the section on descriptive mineralogy, both to change the classification to that based on groups of chemical compounds, now nearly universal, and to make the individual descriptions of species more coherent and adaptable to reference. Discussions of the basic mineralogical concepts, such as the definition of a mineral, have been brought into accord with current ideas. An interesting section on crystal growing has been introduced, enabling the student not only to see for himself how crystals may be formed, but also to possess crystals well enough formed for study, without having to buy expensive specimens from a dealer.

In the opinion of this reviewer the book is open to criticism on only two points of any significance. Firstly, it is doubtful if the brief explanation of polymorphism given would make quite clear to the beginner that the polymorphs differ in atomic structure, though the atoms involved are exactly the same in both kind and number. Secondly, a few of the diagrams in the section on crystallography contain errors which the crystallographer would recognize as such, but which might easily lead to misunderstanding on the part of the impressionable young beginner. An example is figure 45, showing the tetragronal axial cross and the first order tetragonal prism: the a_2 axis fails to intersect the edges of the prism. Such errors may not seem serious, but it might be remembered that this book is intended for beginners and should impress upon them the need for accuracy and precision in advanced study of the subject.

Doctor Hurlburt has produced a commendable revision. The illustrations, with the exceptions noted above, are good, and a very definite improvement over the previous edition is the use of actual photographs to illustrate some common mineral habits. There are few typographical errors, and the general make-up of the book allows easy and enjoyable reading. It is a worthwhile addition to the library of any beginning mineralogist. WALDEN P. PRATT

Animals Without Backbones: *An Introduction to the Invertebrates*; by RALPH BUCHSBAUM. Revised edition. Pp. xii, 405; 160 pls., and numerous figs. Chicago, 1948 (Chicago University Press, $6.50). Intended as an introduction to invertebrate zoology for college students this lively textbook has found wide use at several scholastic levels and has gained a popular following as well. While definitely not a popularization of the subject, the interesting nontechnical style and excellent illustrations put the subject matter within the grasp of the interested layman. As a text it is unusually adaptable because the author has organized the material in such a way that either a general or a detailed survey of the subject can be made without loss of continuity or omission of basic principles.

The text of the revision remains much as it was in the first edition except for the changes necessary to bring it up-to-date and for the addition of a short chapter at the end of the book. This new chapter, entitled *Further Knowledge*, introduces the student to scientific literature, explaining the fields of specialization within the biological sciences and giving their principal media of communication. A selected bibliography of about 70 items bearing on the invertebrates is included and three short scientific articles, reprinted in their original form, serve as examples.

One minor criticism of the text is its lack of any reference list showing, in its entirety, the classification of invertebrates used in the book. The omission is especially noticeable in light of the fact that an entire chapter is devoted to the subject of classification.

The new edition is even more copiously illustrated than its predecessor with thirty-two new pages of excellent gravure illustrations appearing in the photographic sections. By bringing together such an exceptional collection of photographs Buchsbaum has indeed accomplished his stated aim of giving the student a vicarious field experience.

The revised edition of *Animals Without Backbones* will be welcomed by teachers of elementary zoology and paleontology. It is an outstanding general science text and sets a standard that other college texts at this level may well imitate. KARL M. WAAGÉ

Recent Marine Shell-Bearing Mollusca of the South Sea Islands (*Part* 1); by KOTORA HATAI. Bulletin 7., Tropical Industry Institute, Palau, South Sea Islands. Occasional Paper No. 24, Institute of Geology and Paleontology, Tohoku Imperial University, Sendai, Japan, 1941. Published in 1941, this is an admittedly incomplete work, planned as the first in a series on the recent shell-bearing mollusks of the South Sea Islands. The author has covered the larger and more spectacular gastropods and pelecypods quite thoroughly, and subsequent volumes undoubtedly were planned for the smaller and less well-known mollusks. The war, of course, halted all activities of this nature, and it is now problematical whether or not the original project will ever be carried out.

The report is in two volumes: part A consisting of 159 pages of text, and part B of 79 plates. The author gives a brief quotation from the original description for each species, and discusses its known geographic range. Throughout the text the generic names that are familiar to all shell collectors are used. Thus all of the cowries are under the genus *Cypraea,* notwithstanding that the genus has been split into several genera by modern malacologists. However, in the fore part of the book the modern classification of Dautzenberg, Leloup, and others is given. This seems an admirable arrangement, as the non-specialist can find his shells easily by the names they have carried for scores of years, and a quick reference to the modern list will give him the up-to-date name.

The illustrations are generally excellent. Reproduced in collotype, each shell is shown in two or more views, and all but the very large varieties are at natural size. The photography and reproduction are both excellent, but the choice of specimens for illustration was not in every case a happy one. A few characteristically marked species such as *Conus marmoreus* and *Cypraea argus* are illustrated by worn shells that fail to show the mollusks' true beauty. Also, some of the common and well-known shells such as *Cypraea mappa* and *Conus imperialis,* present in nearly every lot of shells from the South Seas, are not illustrated, altho they are mentioned in the text. The lack of an index is disappointing.

Nearly every small city museum and preparatory school, and many high schools and libraries, have boxes of sea shells that have been "donated" at one time or another. Our returning service men and women have probably brought back tons of shells from the South Sea area, and in the years ahead it is likely that more and more "donations" to local institutions will be made. The present work should be extremely useful to the curators of such institutions, as it would be a simple task to get the correct name and approximate locality of almost any of the commoner shells from this area. P. A. MORRIS

Tellus, a quarterly journal of geophysics.—A new journal has been launched by the Svenska Geofysiska Föreningen (Swedish Geophysical Society), as an international forum for original and review articles on geophysics. Volume 1, No. 1, dated February 1949, contains 64 pages and carries six articles and two "notes," all in English. Contributions will be accepted in English, French, or German. The annual subscription rate is U. S. $6.00 or Swedish Kr. 21.00. Contributions and subscription orders may be addressed to the Editor, Tellus, Sveriges Meteorologiska och Hydrologiska Institut, Fridhemsgatan 9, Stockholm, Sweden.

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METASOMATIC GRANITIZATION OF BATHOLITHIC DIMENSIONS

PART III. RELATIONSHIPS OF SYNKINEMATIC AND STATIC GRANITIZATION

PETER MISCH

ABSTRACT. Part III consists of a discussion of the relations between synkinematic granitization as described in Part I, and static granitization as treated in Part II. Three areas in which both occur are briefly described. In the synkinematic gneisses of Nanga Parbat, relatively small aplitic-pegmatitic bodies formed after orogeny had ceased. In the adjacent noritic masses, static granitization produced small quartz-dioritic and granodioritic bodies. At Ts'angshan in West Yunnan, Archaean gneisses show synkinematic granitization, but on a local scale granitization continued after the end of deformation. Continuous structural evolution from synkinematic to static forms of granitization is described. West of Riverside in North Central Washington, Paleozoic rocks have been intensely deformed, variously metamorphosed, and partially granitized. There are all transitions from synkinematic to static granitization. The latter has formed large "eugranitic" masses and in a number of places extended beyond the limits of synkinematic granitization.

Criteria for synkinematic and static granitization are summarized. Genetically, synkinematic granitization is presented as a fundamental process in geosynclinal orogeny, and static granitization as its continuation after the end of orogeny. Postorogenic conditions favor a carrying of the process into higher levels of the crust. Where only static granitization is visible, the products of previous synkinematic granitization may be concealed at depth. Granitic batholiths are considered as predominantly metasomatic, and the rise of magma in broad fronts to shallow levels without breaking through as unlikely. Mobilization of metasomatized rock masses may more locally occur as a still more advanced stage of granitization. To sum up, genetic unity of synkinematic and static granitization in geosynclinal belts is emphasized.

INTRODUCTION

IN Part I synkinematic granitization of batholithic dimensions has been described, and in Part II static granitization on a similar scale has been treated. Emphasis has been on the evidence for regional metasomatism, as against igneous intrusion. The subject of Part III is the genetic relationship of synkinematic and static granitization. Prior to the general treatment of this subject, some examples are given from areas in which synkinematic and static granitization occur together. Emphasis is mainly on the distinction between and the relationships of the synkinematic and the static process. In order to avoid repetition, the evidence for metasomatism as such is only briefly sketched in these areal descriptions, particularly where such evidence is similar to that offered in Parts I and II, and additional data apply only to modifications not treated in Parts I and II. The areas have been selected so as to offer some variety, with regard to time relations, chemical environment, and relative scale of synkinematic and static granitization.

I. NANGA PARBAT AREA, NORTHWEST HIMALAYAS

The first example is the Nanga Parbat region which was described in Part I as a typical case of synkinematic replacement on a vast scale. Later, statically formed granitic rocks are subordinate in this region, and are not as grandly displayed as in some other areas, but this region has been selected as an example because the general background of synkinematic transformation has been established in Part I.

I do not refer to those synkinematic granitic gneisses in wide parts, and especially the interior, of the Nanga Parbat massif, in which crystallization lasted somewhat longer than deformation (cf. Part I). This does not establish a separate phase of static replacement, but merely indicates that temperature was still high in these rocks when orogenic deformation ceased. Not even cross-cutting pegmatites or aplites were formed at this stage, but all mechanical lattice disturbances were obliterated. The statically formed rocks with which we are concerned are distinct bodies of later granitic rocks which lack schistosity and transect the structures of the rocks described in Part I. These bodies comprise two varieties which are associated with country rocks of different chemical composition. They are (a) mostly tourmaline-bearing acidic granites in the gneiss massif, and (b) quartz-diorites and granodiorites in the masses of altered Eocene norites.

(a) *Tourmaline-granites in Nanga Parbat gneiss massif*

The tourmaline-granites occur in discordant bodies, comprising small dikes and large dike-like and lenticular masses which attain a maximum size of a few miles and are thus still very small compared to the huge area of synkinematic granitization in which they are irregularly scattered. The *field relations* of most of the larger masses suggest a replacement origin rather than forceful mechanical intrusion (cf. fig. 1, and pl. 1, fig. 1). Generally no signs of mechanical displacement of the structures of the enclosing gneisses were observed. At several localities thin septa and lenses of gneiss are preserved within the granite without having been disturbed or disrupted. At least some of the small dikes show field features suggestive of replacement (cf. G. E. Goodspeed's 1940 criteria for replacement and dilation dikes). An example is shown in plate 1, figure 2, in which fracture control is also indicated. There is, however, a possibility that some of the dikes and dike-like masses were formed by dilation and actual injection of mobilized material. Further detailed field work would be required to clarify this point.

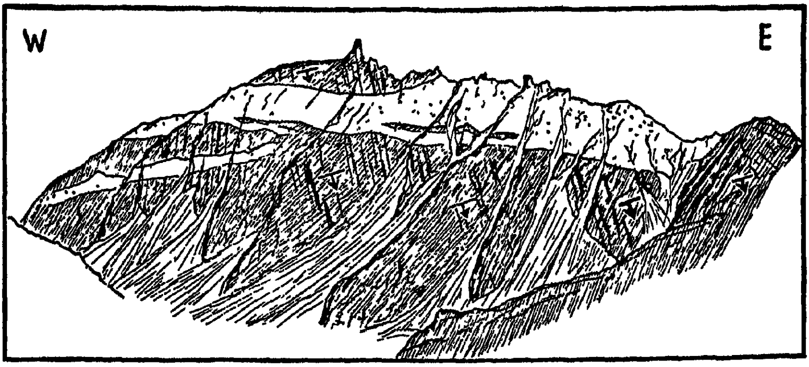

Figure 1. Postorogenic tourmaline-granite in synorogenic granitic biotite-flasergneiss South face of Jabardar Peak (14,800′), 12½ miles north of Nanga Parbat. The face is about 4,000 feet high. After Misch (1935, fig. 23).

The rocks are all very leucocratic. Many have an aplitic, some a pegmatitic character. The main constituents are *quartz*, *potash-feldspar*, and *sodic plagioclase* (partly albite, partly albite plus oligoclase, and rarely oligoclase alone; mostly zoned). In nearly all specimens the quartz-feldspar fabric is *crystalloblastic*. Quartz and feldspar interlock and penetrate each other. Myrmekite is common. Occasionally large porphyroblasts of potash-feldspar enclose the other minerals. More or less recrystallized cataclastic structures sometimes occur. They indicate a continuation, during the crystallization of the rock, of tectonic fracturing which was presumably an important factor in determining the original location of the scattered granitic bodies.

Only one specimen has an igneous-appearing texture, with subhedral feldspars and interstitial quartz, and two or three others combine igneous-appearing and crystalloblastic features. However, "interstitial" textures *can* form by replacement, as was seen in some granitized basic rocks in this area (cf. Ib), though in this particular tourmaline-granite local "rheomorphism," to use H. G. Backlund's term (1937; cf. also G. E. Goodspeed, 1948, 1949), is perhaps a more likely explanation. Action of hot gases may have helped to achieve this result. Their presence is indicated by "pneumatolytic" minerals.

Tourmaline is present in most specimens and varies, often in the same thin section, from irregular replacement shapes to perfect crystals some of which still contain quartz inclusions. Very locally tourmaline replaces biotite. Half of the tourmaline-bearing granites contain *fluorite* which irregularly replaces feldspar. One specimen contains *topaz*. In some of the granites minor amounts of *biotite* occur as ragged grains which give the impression of being absorbed and replaced by felsic minerals. It is often marginally surrounded by later *muscovite* which may also replace feldspar. The biotite in the granites optically resembles that in the enclosing flasergneisses, and I would interpret it as a relic which has survived from the gneissic country rock. At contacts of granite and gneiss a similar absorption of biotite and formation of new muscovite can be observed. To a certain extent the disappearance of biotite is linked with the appearance of tourmaline.

The main mineralogical differences between the granites described and the granitic biotite-flasergneiss in which most of them occur, are the important role "pneumatolytic" minerals play in most of the granites, their very low content or complete lack of biotite, and the usually more sodic character of their plagioclase. A small amount of calcium is bound in the fluorite which in part of the granites replaces some feldspar, but as a whole the granites are poorer in calcium than the enclosing gneisses. The same holds true of iron and magnesium. It thus appears that the mineralogical-chemical differences between the original argillite-derived schists and the granitic biotite-flasergneisses on the one hand, and between these latter and the acidic granites on the other hand, point in the same direction. A special feature in the scattered small granite bodies is concentration of boron and fluorine. An active role of hot gases is obvious, definitely at a late stage, and possibly even during earlier stages of the formation of these rocks.

Exceptionally, local static granitization of lime-silicate-granulites was observed in the interior of the Nanga Parbat massif. One example shows the typical lime-silicate mineral assemblage with an additional large amount of orthoclase the potash of which cannot have been derived entirely from the original argillaceous limestone. A more advanced stage of the same process seems to be represented by a small body of pyroxene-granite which consists of a large quantity of orthoclase and smaller amounts of plagioclase (An 31+), quartz, diopsidic pyroxene, and sphene. During a number of traverses of the gneiss massif only one single small outcrop of a granitic rock free from potash feldspar was met (a leucocratic biotite-quartz-diorite).

(b) *Quartz-dioritic and granodioritic rocks in the metanorite masses*

In contrast to the granites in the gneiss massif, *statically formed quartz-dioritic, trondhjemitic and granodioritic rocks* occur in the large metanorite masses bordering both sides of the gneiss massif (cf. fig. 1, in Part I). At the time of the metamorphic recrystallization of the basic masses, deformation was restricted to their margins, and to local zones in their interior, mostly near included metasediments. Otherwise, the metanorites are non-schistose. They include some finer-grained hornfelsic amphibolites part of which may be of sedimentary derivation. The synkinematic granitization in the Nanga Parbat massif has to some extent also affected the norite-derived amphibolites in the border zone where they are interfolded with the gneisses. Within the metanorite masses, static granitization has been active at many places though most of the individual transformed bodies are very small. This phase of static granitization appears to be later than the formation of the schistose amphibolites and of the synkinematic gneisses, for where it has affected schistose varieties of the basic rocks, it is later than the schistosity. On the other hand, it is probably somewhat earlier than the formation of the tourmaline-granites in the gneiss massif.

The small statically granitized bodies in the metanorite masses belong to the "eugranitic-appearing type" mentioned at the end of Part II. They form light-colored pockets and patches which spread out irregularly in the basic rocks and locally pass into replacement breccias where fracturing has occurred (pl. 2, fig. 1). There are also replacement veins and dikes. Elongated schlieren-like patches may occur where the country rock is weakly schistose (pl. 1, fig. 3). Contacts are locally sharp, especially where fracture control is visible, but are mostly gradational, and intermediate rock types are common. In the locally occurring schistose varieties of the basic rocks, weak relict schistosity has sometimes survived in granitized portions. That granitization was postkinematic, is shown in such cases by irregularly crosscutting contacts, and by a transverse orientation of some of the feldspar porphyroblasts (cf. below, discussion on crosscutting dikes with relict schistosity at Ts'angshan).

The granitized rocks are enriched in plagioclase and quartz, impoverished in mafics, much hornblende has been replaced by biotite, and some potash feldspar has formed. Muscovite occurs locally. The plagioclase has become more sodic than in the basic rocks (except where their plagioclase had, during recrystallization prior to granitization, already been completely broken up into epidote and oligoclase). The hornblende of the metanorites and hornfelsic amphibolites survives in some of the granitized rocks, whereas in others part or all of it is biotitized and potash-feldspar may also have formed. At contacts, biotitization may occur in the basic rocks.

All the granitized rocks are crystalloblastic. Sutured granoblastic textures are very common. Poikiloblastic growth, similar to that in the hornblende of many metanorites, is often seen in the granitized rocks. Plagioclase often forms replacement porphyroblasts enclosing small relict grains. Where potash-feldspar has formed in larger quantities, it mostly consists of similar replacement porphyroblasts. At several places, a peculiar recrystallization texture was noticed in relatively fine-grained intermediate rocks. Their plagioclase forms a xenoblastic polygonal pavement which may persist when the rock is enriched in plagioclase, but with further progress of granitization gives way to more coarse-grained sutured-granoblastic textures.

Quartz occurring between crystalloblastic feldspars sometimes has the appearance of an "interstitial filling." Recrystallized mafics occasionally occupy a similar "interstitial" position. Generally textures are uneven. Hornblende and biotite, partly associated with epidote, are commonly concentrated in clusters which are recrystallized relics of the basic country rock and often grade into the latter. There are all passages from feldspathized rocks which still contain much of the fabric of the basic rocks between plagioclase porphyroblasts, to leucocratic rocks in which only minor amounts of unevenly scattered mafics have survived. Where relict schistosity occurs, it is usually marked by the mafics.

Some of the small granitized bodies bear evidence of local metamorphic differentiation, but in most cases an addition of alkali and silica from a more distant source is indicated by the composition of the original and the metasomatized rocks. Sodium is commonly the main alkali added, though part of the total sodium present was contained in the original rocks. Potash-feldspar is often absent, and only locally has it formed in larger quantities. Potash metasomatism is frequently confined to biotitization, and even this may be lacking. Where metasomatic potash-feldspar is present, it has mostly formed at a late stage of transformation.

The *chemical contrast between the granitic rocks in the gneiss massif, and the predominantly quartz-dioritic to granodioritic patches and veins in the metanorite masses*, is striking. It reflects the different chemical composition of the original rocks in both areas. This regional relationship is a strong additional argument in favor of granitization. The different sodium-potash ratio is in part directly due to the different composition of the country rocks, and in part to a selective character of metasomatism, basic rocks being primarily in need of sodium (and silica), and usually being able to absorb larger quantities of potash only at a later stage of their transformation (cf. Part I, pp. 237-238). Moreover, where metamorphic differentiation within the metanorites has contributed the main alkali transferred was, of course, sodium.

Larger bodies of "eugranitic" rocks which extend over several miles, were observed near the eastern border of the area, both in the metanorites and the Cretaceous-Eocene volcanic formation. Their composition shows the same variation as that of the small granitized patches and veins, but as a whole rocks which, in addition to plagioclase, contain considerable amounts of potash-feldspar, are more common. Limited time did not permit collecting sufficient data on these rocks to warrant a general genetic interpretation. However, such data as were obtained, show that at some places metasomatic processes were active in the formation of these rocks.

The specimens obtained have coarse-grained granoblastic fabrics. Hornblende and biotite are unevenly scattered and often concentrated in clusters which suggest recrystallized relics of the basic country rocks. Poikiloblastic growth occurs. A granodiorite which covers a large area encloses, in a granoblastic plagioclase-quartz-hornblende fabric, large microperthite idioblasts which replace the granoblastic matrix and enclose numerous relics of hornblende, plagioclase and quartz. The porphyroblasts reach an inch across and are usually Carlsbad twins. Introduction of potash was the latest stage in the development of this rock. The granodiorite encloses more fine-grained relics of incompletely feldspathized hornfelsic amphibolites. As their transformation progresses, only clusters of more coarsely recrystallized hornblende remain.

East of Kalapani (cf. fig. 1 in Part I) a "eugranitic" rock mass is in contact with Eocene-Cretaceous basic lava. Near the contact, the basic rock has been statically recrystallized to a fine-grained compact hornfelsic amphibolite. Where fracturing had occurred in the basic rock, replacement breccias have formed at the contact. Within the breccia, contacts are partly sharp where formed by fractures, but there are also all gradations from the basic into the acidic rock. Plagioclase porphyroblasts developed in the basic rock, and the hornblende was more coarsely recrystallized. As feldspathization advanced, the hornblende became restricted to poikiloblastic relict patches and clusters. Quartz was introduced, and the rock became strongly leucocratic. At a comparatively early stage of transformation, tiny patches in the rock occasionally developed certain igneous-appearing microscopic features; plagioclase porphyroblasts developed crystal form, quartz grew between them, and thus an interstitial-appearing texture was produced. If such texture is found in a more fully granitized rock, as for instance some miles farther north, it might be taken as evidence of an igneous origin, but obviously it can also form by metamorphic-metasomatic processes.

II. TS'ANGSHAN AREA, WEST YUNNAN (CHINA)[1]

(a) *Summary of synkinematic granitization in the Ts'angshan gneisses*

North of the Burma Road and west of the ancient town of Tali, the Ts'angshan range extends in a north-northwesterly direction, rising to over 14,000 feet. This region is near the eastern border of the geosyncline which I found to occupy West Yunnan, and is characterized by overthrusting to the east (Misch, 1945a and 1945b). One of the thrust masses consists of Archaean gneisses. It composes most of the Ts'angshan range. Near the thrusts bordering it below and above, this mass has suffered retrogressive dynamic metamorphism which was contemporaneous with progressive dynamic metamorphism[2] in Paleozoic and Mesozoic rocks. In the interior of the mass, the gneisses have remained unaffected. It is these rocks with which we are concerned. They originated from a thick succession of partly sandy argillites with subordinate basic layers. These rocks have been intensely folded and during this orogeny have been metamorphosed into fine-grained biotite-paragneisses and biotite-schists, quartzitic biotite-schists, and some more hornfelsic schists, with subordinate amphibolites (partly para-amphibolites grading into biotite-schists). Still during orogeny, these metamorphics have been variously granitized.

The metamorphic migmatites thus formed show, in field sections, hand specimens, and thin sections, a very close resemblance to the various gneisses of Nanga Parbat described in Part I which are representative of synkinematically granitized argillites the world over. The Ts'angshan gneisses comprise the same types of feldspar-porphyroblast-schists, augengneisses and banded gneisses which occur at Nanga Parbat. There are the same gradations between and the same interlayering of these types, the same variations in the proportion of non-replaced and of metasomatized material, and in the thickness of individual bands. The banded gneisses show the same evidence of having been formed chiefly by *lit-par-lit* replacement rather than true injection. There are also similar granitic biotite-flasergneisses. They have here developed from augengneisses. There are also zones of white microcline-gneisses which resemble certain rocks of Nanga Parbat. The schists and gneisses of Ts'angshan are essentially mesozonal.

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[1] My field work in the Ts'angshan area was done during a number of trips from 1939 to 1944 under the auspices of National Sun Yat-sen University, National University of Peking, Geological Survey of Yunnan, and a Committee for scientific exploration of the Tali area formed by Tali citizens and headed by General Tsung-Liu Ma. A report for publication in Chinese was submitted. I hope soon to publish the map and English report.

[2] I use the term "dynamic metamorphism" to indicate such low grade (essentially epizonal) metamorphism which can be demonstrated to have been caused by intense mechanical rock deformation (differential shearing, etc.) without any appreciable contribution by other factors (introduction of heat, metasomatism, etc.).

The close petrographic similarity of the Ts'angshan and Nanga Parbat rocks and the presence of the same replacement features prove an identical origin. I may therefore refer to the genetic discussion in Part I. The main difference is that at Nanga Parbat the zones of progressive metamorphism and granitization are displayed in unusual completeness and regularity, whereas in the much older[3] rocks of Ts'angshan only part of the picture has been preserved, owing to post-Archaean erosion and to later folding and thrust-faulting.

(b) *Local postkinematic granitization*

Granitization in the Ts'angshan gneisses was chiefly synkinematic as is shown by sharp parallelism of bands, alignment and elongation of porphyroblasts, well-marked schistosity, and other megascopic and microscopic structural features. However, subsequent static replacement was widespread, although it occurred on a small scale and did not lead to the formation of any larger bodies of postkinematic granitic rocks within the gneiss mass.

Well-exposed examples are easily accessible near the Burma Road in a canyon which cuts across the southernmost part of the Ts'angshan range. Figure 2 shows a small pegmatite dike from this locality. On the one hand, it transects the schistosity of the country rock and is therefore later than the differential rock deformation, i.e. postkinematic. On the other hand, marked chiefly by coarsely crystallized biotite, it still possesses the schistosity of the country rock, and cannot therefore have

[3] Referring to the age of metamorphism and granitization, not of the original sediments, in both areas.

formed by postkinematic igneous intrusion cutting across the schistosity of the country rock. The obvious explanation is that the dike is of replacement origin. It is crosscutting because it is postkinematic, and it has inherited the schistosity of the country rock as a relict structure because it is metasomatic.

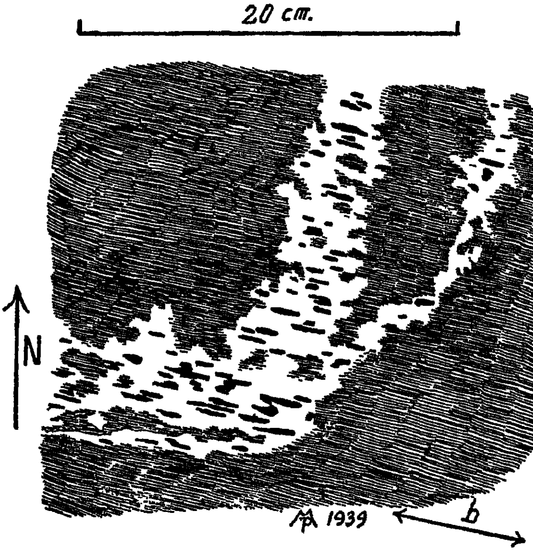

Figure 2. Postkinematic replacement pegmatite in fine-grained schistose biotite-gneiss. Erho Canyon, Ts'angshan, West Yunnan. Relics of schist in the pegmatite are shown in gray, coarsely recrystallized biotite in black b = lineation.

This is confirmed by the crenulated border of the dike, by the presence of large feldspar porphyroblasts invading the country rock along the contacts, and by the preservation of undisturbed (neither shifted nor rotated) relics of the fine-grained schistose country rock within the dike ("skialiths," to use G. E. Goodspeed's 1948 term). Small pegmatite bodies of similar origin often occur in the Ts'angshan gneiss. Some of them have lost their relict schistosity, and may in some instances have actually moved.

Another example of static granitization is shown in figure 3. In this example also the metasomatized rock possesses relict schistosity, and has postkinematic contacts which irregularly transect the schistosity of the country rock. Contacts are partly gradational and partly sharp. It has been emphasized in an earlier chapter that granitization is capable of forming sharp contacts, especially if there is joint or fracture control. Figure 3 shows, in addition to pegmatite, a more fine-grained, light-colored, somewhat aplitic granite with relict schistosity. Small bodies of this type of replacement rock are quite common in this area. Some of them have lost their relict schistosity and become directionless granite. Further west in the canyon, outcrops of this rock are well-exposed on the Burma Road, and contain completely undisturbed relics of paragneiss with feldspar porphyroblasts and granitized bands. These relics are excellent examples of what G. E. Goodspeed has termed "skialiths." Small veins of postkinematic replacement aplite are also often found in the Ts'angshan gneiss.

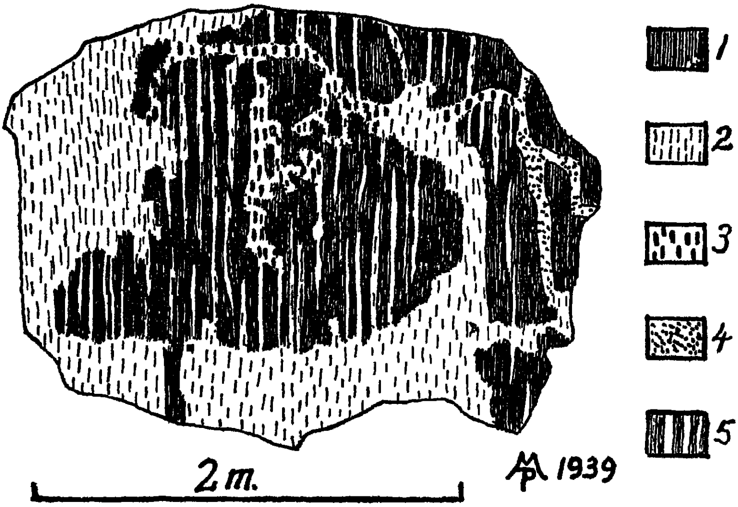

Figure 3. Postkinematic replacement in gneiss. Erho Canyon, Ts'angshan, West Yunnan.

1 = fine-grained schistose biotite-gneiss
2 = leucocratic granite } postkinematic replacement rocks, with gneissic
3 = pegmatite } relict schistosity
4 = quartz veins
5 = banded gneiss, formed by postkinematic mimetic *lit-par-lit* replacement

The local postkinematic granitization in the Ts'angshan gneisses is of particular interest because it is continuous with the synkinematic replacement in these rocks. The process of metasomatism had reached a climax during the later part of the period of orogenic deformation, and it still continued, although on a more local scale, when deformation ceased. It is therefore possible to compare the effects which the same granitizing solutions have had on the same rocks under different mechanical conditions, i.e. (1) during differential deformation of the country rock, (2) at the time of waning deformation, and (3) after deformation had definitely ceased.

As to (1), the metamorphic migmatites thus formed show sharp parallelism of all their structures and well-marked *schistosity* which was *mechanically active* at the time of formation of these rocks. This "active schistosity" differs from *passively inherited relict schistosity* in its appearance as well as in the behavior of the component mineral grains (kind and degree of preferred orientation, combination of growth and deformation features). There is no need to illustrate examples of synkinematically metasomatized Ts'angshan gneisses, since their appearance is identical with that of the synkinematic gneisses of Nanga Parbat, and I may therefore refer to the illustrations in Part I.

Rocks metasomatized at the time of waning deformation are transitional in character. An example is shown in plate 2, figure 2. Parallelism of banding, and schistosity in the transformed layers, are less sharp than in rocks which were replaced under fully synkinematic conditions. A certain tendency of the metasomatizing solutions to spread in directions other than that of the schistosity of the country rock, becomes for the first time visible but still remains largely repressed. It might be objected that the rock shown in plate 2, figure 2, was granitized under entirely static conditions, that its banding was guided by the pre-existing schistosity of the country rock, and that the schistosity of the replaced portions is an inherited relict structure—a process which may be called "mimetic replacement." This process has undoubtedly taken place in many metamorphic migmatites, but I have nowhere seen rocks with mimetic replacement in which part of the metasomatizing solutions have not succeeded in "emancipating" themselves from guidance by the mechanically inactive older schistosity of the country rock, and in thus forming some cross-cutting contacts which interrupt the parallelism of mimetic replacement.

Figure 3 in plate 2 shows latest-kinematic replacement succeeded by early post-kinematic metasomatism. The former is similar to that shown in plate 2, figure 2, and the latter has formed small cross-cutting pegmatitic-aplitic veins. Their replacement origin is shown by gradations and by some inherited schistosity which is chiefly marked by small relics of biotite-gneiss the biotite of which has usually been more coarsely recrystallized. The vein on the left still shows, on a very small scale, some mimetic guidance of replacement by the schistosity of the country rock, but as a whole this vein transects the schistosity. The continuous transition from syn- to post-kinematic metasomatism is obvious in this rock. A more advanced stage of post-kinematic replacement is illustrated by figure 4 in plate 2. Leucocratic granitic veins transect medium-grained biotite-flasergneiss in various directions, and are only very locally (center of figure) guided by its schistosity.

Small statically formed metasomatic veins and patches of the types shown in text figures 2 and 3 and in plate 2, figures 3 and 4, are common in the Ts'angshan gneisses. They are typical of those found in many other areas in which granitization has continued after the end of orogenic deformation. Similar patterns are frequently developed on a much larger scale. In fact, the detailed features which have been figured, are *diminutive models of patterns displayed on a regional scale in many granitized areas.*

III. RIVERSIDE-CONCONULLY AREA, NORTH CENTRAL WASHINGTON[4]

Some observations will be briefly presented which have a bearing on the questions treated in the present paper. This preliminary note will be followed by a fuller description elsewhere.

(a) *Outline of areal geology*

In the Riverside-Conconully area west of the Okanogan River, a thick geosynclinal Paleozoic succession has been intensely folded, variously metamorphosed, and in part granitized, presumably in Jurassic time. The lower part of the column consists of dolomites and limestones most of which have been altered to banded marbles. They conformably underlie rocks equivalent to parts of the "Anarchist series" established by R. A. Daly (1912) near the border and extended to the south by A. C. Waters and K. Krauskopf (1941). Fossils locally preserved in the limestone indicate a Paleozoic and, according to a preliminary examination by H. E. Wheeler, probable Devonian, age. The upper part of the column consists chiefly of thick argillites altered to phyllites and biotite-schists, with actinolite-tremolite-schists and granulites in their lower part. Most of these latter were derived from impure dolomitic sediments, and grade laterally into amphibolites. Synorogenic regional metamorphism of varying degrees has been succeeded in wide parts by static thermal recrystallization. Typical examples are phyllitic biotite-schists containing large post-kinematic andalusite porphyroblasts. Many of the originally synkinematic schists show superposed hornfelsic textures.

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[4] I studied this area when conducting the 1948 geologic field course of the University of Washington. The area is located southwest of the region in which Waters and Krauskopf have described the "Anarchist series" and the "Colville batholith" (1941). I wish to thank Aaron C. Waters for introducing me to this region during a joint field trip, in which Richard W. Lounsbury also participated. The latter assisted me during the field course; his Stanford Ph.D. thesis will deal with an area farther north.

The degree of metamorphism is clearly related to granitization which comprises a synkinematic and a postkinematic phase and thus matches the history of metamorphism. Granitized areas covering hundreds of square miles occur to the west and southwest. They contain relict zones of Paleozoic metamorphic rocks in which the predominantly northwesterly regional orogenic trend has been left undisturbed. These metamorphic rocks were still found in the high granitic Tiffany range[5] on the west which forms the divide between the Okanogan and the Methow valleys. A smaller granitized body, measuring about ten square miles, occurs within the sedimentary area west of Riverside.[6] It occupies the core of a syncline, replacing biotite-schists and amphibolitic rocks, whereas the underlying marbles which form the limbs of the syncline, have escaped granitization. Generally, in the Riverside-Conconully region, *the calcareous-dolomitic anticlines have not been granitized,*[7] *and granitization has been restricted to parts of the synclinal areas composed of the younger rocks which were chemically more susceptible to transformation.*

[5] A detailed study of this area is being carried out by R. Goldsmith for an advanced degree in geology at the University of Washington, whereas J. Robertson is doing a Master's thesis on the rocks near Conconully.

[6] A detailed study of the rocks in this granitized body is being made by L. T. Grose for a Master's degree at the University of Washington.

[7] In an area farther south, west of Omak, granites of mostly still undetermined origin contain marble remnants with high-temperature lime-silicates. This area is being studied by W C. Boulton for an advanced degree.

(b) *Summary of synkinematic granitization*

Synkinematic metamorphic migmatites are developed in the western and southwestern portion of the area. In wide parts they have been subjected to subsequent static replacement and as a result they have partly lost their identity. The synkinematic gneisses closely resemble those described in Part I. There are feldspar-porphyroblast-gneisses, banded gneisses, augengneisses and flasergneisses. They have formed from biotite-schists and partly biotitized amphibolites, which often occur as relict bands in the gneisses, and grade into them. Metasomatic feldspar porphyroblasts exhibit the usual replacement features. Structural parallelism is general in the synkinematic gneisses.

Massive granitic-appearing biotite-augengneisses are for instance typically developed south-southeast of Conconully. They occur near the border of the granitic area, and locally include relics of sillimanite-schist. Farther west, a complete transition from lime-silicate-schist into banded gneiss and flasergneiss was observed, with intervening stages of amphibolite and feldspathized amphibolite. In the high Tiffany range on the northwest wide areas of synkinematic migmatitic gneisses with relict bands of biotite-schists and amphibolites were found to occur as far as 11 miles west of the border of the granitic mass forming this range. They demonstrate that metasomatic replacement was not restricted to the marginal parts of this mass, but was responsible for the formation of its central portion as well, i.e. that granitization was of batholithic dimension.[8] Some amphibolites run from the non-granitized eastern region into the western granitic area and can be traced to the summit of the high Tiffany range.

Most of the synkinematically granitized rocks have textures which indicate that their crystallization lasted longer than their deformation. In view of the widespread postkinematic continuation of granitization in this region, it is not surprising that even those rocks which were not directly affected by the postkinematic phase of replacement, generally remained hot enough to continue their recrystallization for some time.

(c) *Summary of postkinematic granitization*

The static phase of granitization in this area is characterized by crosscutting contacts and by predominantly direction-

[8] I have seen similar conditions east of the Okanogan Valley.

less granular fabrics. At many places, it has transgressed the limits of synkinematic granitization. We may therefore distinguish between static replacement (1) in synkinematic gneisses, and (2) in previously non-metasomatized metamorphics. On the basis of the size of individual replacement bodies, we may differentiate into static granitization (a) restricted to small-scale features, and (b) responsible for the formation of larger "eugranitic" bodies.

Case (1a) shows impregnation of synkinematic gneisses with small replacement veins and patches which have crosscutting and irregular contacts and sometimes have preserved relict schistosity. These small replacement bodies closely resemble those described above from Ts'angshan, and there is the same continuous structural evolution from synkinematic to postkinematic forms of replacement. I refer to the description in chapter II, and to figures 2-4 in plate 2 all of which might have been photographed in the present area. These features are, for instance, typically developed in part of the Tiffany range on the west of the area. Here metasomatic formation of postkinematic pegmatites is well displayed. Large pink potash-feldspar idioblasts began to grow in the gneiss, concentrated in small groups, and finally coalesced with the coarsely recrystallized remnants of the gneiss into patches and veins of pegmatite.

Case (2a), i.e. small-scale static granitization in previously non-granitized rocks, is represented by chiefly aplitic irregular veins and patches. There are all gradations into the country rocks which are mainly biotite-schists and amphibolites, the latter frequently biotitized. As a whole, preservation of relict schistosity is rare. Part of the enclosing schists have a superposed hornfelsic texture. Relics of schist are common in the small replacement bodies, and do not show any signs of disturbance, not even where they form thin delicate lenticles and septa which have preserved the direction of schistosity of the country rock within replacement dikes. Good examples are exposed northeast of Conconully. Here and at other localities, also occur true dilation dikes[9] of aplite. I would attribute them to mobilization and intrusive movement of granitized material, rather than to intrusion of original granitic magma.[10] At the locality northeast of Conconully, even barely granitized biotite-schist has, in the form of a cross-cutting dilation dike, intruded biotite-schist.

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[9] As to criteria for "dilation dikes" and "replacement dikes," I refer to G. E. Goodspeed (1940).

Apart from the small-scale and incipient features described, static granitization has in this area produced larger bodies of "eugranitic" rocks, in areas of synkinematic gneisses as well as of previously not granitized metamorphics (cases 1b and 2b). The rocks are predominantly dioritic, quartz-dioritic, trondjhemitic and granodioritic though granitic varieties also occur.[11] These rocks usually lack relict schistosity. Their texture is crystalloblastic and frequently porphyroblastic. Most of the rocks have an even-granular overall appearance, others look porphyritic, often owing to the presence of late potash-feldspar porphyroblasts. Contacts are mostly, but not always, transitional. More fine-grained and basic relics of country rocks are extremely common and show all stages of transformation. In spite of their mostly rather uniform overall appearance, the rocks generally look patchy and uneven in detail where well exposed, often with slightly darker shadow-like areas which to some extent are due to variations in texture, but chiefly in composition, especially in the amount of mafics. I would attribute all these heterogeneities to replacement origin. Microscopic clusters of mafics suggest recrystallized remnants of country rock.

The features described occur in the western areas, for instance in a large chiefly quartz-dioritic and tronhjemitic mass on the eastern slope of the high Tiffany range, as well as farther east, beyond the eastern border of the region affected by synkinematic granitization. North of Conconully near the eastern border of the main area of granitization, relict inclusions, replacement breccias, and granitized patches resembling the "small granodioritic blocks" described by G. E. Goodspeed (1937) from Oregon, occur together, with all transitions from the country rocks through porphyroblastic to "eugranitic" rocks. Here and at several other places, schistose and more compact amphibolites have been successively transformed into gabbroid, dioritic, and quartz-dioritic rocks; where potash introduction has led to the formation of frequently porphyroblastic potash-feldspar, the granodioritic stage has been reached.

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[10] References to the ideas of several authors on "diapirites," "migma," "rheomorphism," "neomagma." are given below in chapter IV

[11] Southeast of the region here described, a large area of leucocratic granite occurs near Omak. Its origin is still undetermined.

Smaller granitized bodies occur still farther east in the metamorphic Paleozoic rocks, and a somewhat larger quartz-dioritic and granodioritic mass occupies a synclinal core west of Riverside (cf. above). It exhibits the replacement features mentioned above. Some relict lenses of amphibolite and biotite-schist can still be correlated to the normal stratigraphic sequence in the syncline. The amphibolite is a more highly metamorphic facies of the actinolite-tremolite rocks mentioned above, and grades into gabbroid-dioritic rocks. Apart from the directionless granular rocks, gneissic rocks occur more locally. Most of them are augengneisses containing varying amounts of potash-feldspar porphyroblasts. They grade into directionless porphyritic-appearing granodioritic rocks with the same porphyroblasts. Apart from the nature of the granitized rocks and their relations to the metamorphic sediments, the fact that the calcareous-dolomitic limbs of the syncline have not been affected, in itself suggests metasomatic replacement of selective character.

Whether the larger statically granitized bodies in this region have to some extent been affected by "rheomorphic" processes (cf. below, chapter IV), is difficult to decide without further detailed study. However, most of these rocks have not undergone mass movement, but remain where they were formed through metasomatic processes.

IV. DISCUSSION OF GENETIC RELATIONSHIPS OF SYNKINEMATIC AND STATIC GRANITIZATION[12]

(a) *Criteria for synkinematic and static crystallization and granitization*

Since sometimes *all* granitization is treated as a static process, it appears appropriate to summarize briefly the *criteria for synkinematic crystallization and granitization* of the type described in Part I. The higher grades of regional metamorphism in geosynclines were in Part I found to be intimately linked with granitization and to depend on it for a supply of additional heat. Therefore, evidence of synkinematic crystallization in higher grade metamorphic rocks is also evidence of a synkinematic character of metasomatism in associated granitized rocks. In fact, in non-granitized and granitized rocks occurring together, crystallization differs only insofar as in the latter, in addition to substances of the original rocks, introduced material has participated. Therefore, both groups of rocks show essentially similar structural features indicative of synkinematic origin. Main field criteria are: well-developed schistosity and a general sharp parallelism of such elements as bands, stringers of porphyroblasts, etc., and an absence of crosscutting relationships—features which are typically developed at Nanga Parbat as well as in many other areas.

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[12] The general discussion in this chapter is based on the rocks described in Parts I and II, and in the first three chapters of Part III. In order to avoid tedious repetition, reference to those descriptions is made only in a few cases.

—Microscopic criteria are numerous. Many of them have been described for regionally metamorphosed rocks in B. Sander's fundamental treatise on metamorphic fabrics (1930, pp. 262-275). Similar features are often found in synkinematically granitized rocks like those of Nanga Parbat. General criteria for synkinematic crystallization include: "snow ball garnets" and other porphyroblasts rotated during growth; the occurrence, in the same thin section, of folds composed of deformed minerals such as micas, and of other folds consisting of undeformed micas in polygonal arrangement ("polygonal arcs"); the presence of polygonally arranged, undeformed, and of folded micas in different parts of the same microscopic fold; the superposition of slight folding in a polygonal arc, both mechanical bending and polygonal arrangement representing the same pattern of deformation; fracturing and subsequent healing of crystals which may result in elongation parallel to the schistosity; generally speaking, the frequent occurrence of several overlapping successive submaxima both of deformation and crystallization, which can be individually dated by certain mineral species (one species sometimes having more than one maximum of crystallization); thus detailed time analysis is possible, and, as a whole, one continuous major process consisting both of deformation and crystallization is indicated. Furthermore, the common metamorphic structure termed "crystallization foliation" (well-developed schistosity without mechanical disturbance of crystal lattices) is rarely due to entirely static mimetic crystallization; under static conditions usually some transversely oriented crystals develop, and if mimetic granitization is added, invariably cross-cutting contacts also form. Typical crystallization foliation may rather be considered as the result of crystallization which was mainly synkinematic and was able to heal lattice disturbances caused by contemporaneous deformation, and which usually lasted somewhat longer than deformation so that any remaining lattice disturbances were eliminated. Additional criteria common in synkinematically granitized rocks are elongation and parallel alignment of metasomatically formed feldspar porphyroblasts and parallelism of replacement bands, apart from frequent traces of mechanical deformation in porphyroblasts which may or may not have been healed by continuing crystallization.

Static crystallization and granitization are free from contemporaneous deformation of the rocks involved. The term "postkinematic" refers to a static process succeeding a phase of differential deformation which may also have been a phase of synkinematic granitization. Except at great depths, static recrystallization of sediments requires introduction of heat,[13] and as I have tried to show in Part II this heat in areas of static granitization is transported by the granitizing agent. Accordingly, the thermal recrystallization of sedimentary areas adjacent to granitized bodies is traditionally included with contact metamorphism.[14] Static crystallization in nongranitized and granitized rocks differs only insofar as in the latter introduced material participates in addition to the substances of the country rock. Both groups of rocks typically have a directionless fabric, lacking schistosity unless it has been inherited as a relic. Previously unmetamorphosed sediments, if thoroughly recrystallized, develop hornfelsic fabrics, with mineral associations varying according to temperature. Bedding is frequently preserved, in the form of differently colored layers differing in composition, and sometimes of division planes. Contrary to some statements in the literature, I have not seen a single case in which bedding in statically recrystallized sediments has become schistosity. The fundamental contrast between a hornfels and a schist[15] illustrates this point. Crystalline schists are not made without differential rock deformation.

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[13] This is confirmed by the preservation in many geosynclines of completely unmetamorphosed sediments five to ten miles thick and sometimes even thicker.

[14] I would restrict the term *contact metamorphism* to thermal metamorphism at true igneous contacts.

Statically granitized rocks are directionless[16] unless they have inherited relict schistosity. In granitized sediments, bedding sometimes survives as a usually faint relict structure, but (contrary to some statements in the literature) it does not become gneissic structure—all products of static granitization of non-schistose sediments I have seen, have directionless fabrics. Characteristic of static granitization are irregular and crosscutting contacts which transect the bedding or the schistosity of the country rock. Granitized patches in the country rock and relics of the latter in the granitized rock, show irregular shapes and are irregularly scattered. Statically granitized rocks and statically recrystallized sediments both have crystalloblastic textures and both lack structures indicative of contemporaneous differential deformation, but the granitized rocks contain metasomatically formed porphyroblasts and have a larger average grain size. Transitional textures are common. Textural and mineralogical relics of the country rocks are widely preserved in the granitized rocks (cf. Part II). At advanced stages of transformation, metamorphic features, such as porphyroblastic textures, usually become less conspicuous, and the textures more igneous-appearing.

Where the rocks subjected to static granitization have been low-grade crystalline schists (phyllites, greenschists, etc.), and, near granitized bodies, have undergone static recrystallization at medium or high temperatures, they tend to approach hornfelsic fabrics, rather than to become typical medium- or high-grade crystalline schists. Lack of stress and deformation during recrystallization may also have its mineralogical expression, as in the synkinematic biotite-phyllites west of Riverside (chapter III) in which andalusite formed when static granitization took place in the vicinity. In granitized rocks of this group, little of the schistosity of the original low-grade schists survives; the rocks are mostly directionless.

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[15] In addition to the differences in structures and textures, there are a few mineralogical differences Kyanite is synkinematic, andulasite static; almandite is much more common in synkinematic rocks, etc

[16] Possible later flow structure due to mass movement of metasomatized rocks, e.g. "rheomorphism" as defined by G. E. Goodspeed (1949), belongs in a different genetic category.

Where the country rocks are medium- and high-grade crystalline schists and synkinematically metasomatized gneisses, i.e. rocks already adapted to higher temperatures, in most cases little change is visible near statically granitized bodies, though there may be a certain tendency to superpose, on the schistose fabric, some hornfelsic or directionless-granulitic features (chapter III). This tendency is often recognizable in thin sections only. In statically granitized rocks of this derivation, the schistosity of the country rocks frequently survives as a relict structure ("mimetic granitization"), but it usually becomes increasingly faint as the process of static metasomatism progresses. Finally, it is lost, and the granitized rock becomes directionless. In other cases, the schistosity of the country rock is lost during an early stage of static metasomatism. As a whole, small replacement bodies more often retain relict schistosity than large ones. The best criterion for a static origin of replacement bodies retaining relict schistosity, is irregular and crosscutting contacts. Microscopic criteria include absence of structures indicative of active deformation, and lack of directional control in the spreading-out of metasomatic feldspar porphyroblasts.

In most cases where synkinematic is succeeded by static granitization, both are merely successive phases in a continuous process. Examples have been described in chapters II and III. Relations between these phases are characterized by transitions, and the contrasting critical features discussed above gradually replace each other. The relative extent and degree of each of the two phases of granitization varies in different areas. The postkinematic phase may be restricted to small-scale features (chapter II, Ts'angshan). Or it may form large bodies and even spread far beyond the limits of the synkinematic phase (chapter III, Riverside). In some cases, the static phase does not occur in immediate continuity with the synkinematic phase, but seems to represent a renewed upward surge of the granitizing agent after it had already subsided. Examples are the postorogenic granitic bodies in the Nanga Parbat gneisses (chapter Ia), in which, incidentally, no relict schistosity was observed.

Can the static processes, *mimetic crystallization* and *mimetic* *granitization*, make rocks which are indistinguishable from those of synkinematic origin? First, it must be realized that crystallization described as "post-tectonic" is in numerous cases crystallization which was chiefly syntectonic and merely lasted somewhat longer than deformation. Examples are strong crystallization foliation as interpreted above, or, to mention a specific feature, such polygonal arcs which must be interpreted in a similar way on the strength of other structural features in the same rock. An example among granitized rocks is the gneiss in the interior of the Nanga Parbat massif which bears evidence of synkinematic origin though its crystallization lasted somewhat longer than its deformation—which, significantly, did not lead to the formation of any crosscutting replacement veins. Contrasted to all these cases is mimetic crystallization as an entirely postkinematic or static process. Its products do not possess the same sharp structural parallelism as those of synkinematic crystallization. Even if a majority of the grains are guided in their growth by the inherited schistosity, yet there are usually others which ignore it, namely transverse crystals and porphyroblasts. The same applies to an altogether static mimetic granitization. Relict schistosity may be preserved, but it becomes successively fainter during the process. It is in most cases marked by inherited mafics whereas the feldspar porphyroblasts tend to ignore it. Contacts of granitized portions may partly follow the pre-existent schistosity of the country rock, but crosscutting and irregular patchy contacts invariably *also* develop (text figs. 2 and 3; pl. 2, fig. 3).

(b) *Genetic relationships of synkinematic and static granitization in geosynclinal belts*

It is generally agreed that the *geosynclinal type of orogeny*[17] is characterized by intense folding frequently accompanied by large-scale thrusting. Equally characteristic of geosynclinal belts of first order (eugeosynclines[18]) are *regional metamorphism*

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[17] The term "geosyncline" is used in its restricted sense ("true geosynclines" of H. Stille, 1924; "mobile belts" of W. H. Bucher, 1933; "orthogeosynclines" of Stille, 1940, and of M Kay, 1944), excluding basins in more stable regions ("parageosynclines" of Stille, 1940; "monogeosynclines" etc. of Kay, 1944). "Orogeny of geosynclinal type" then is equivalent to Stille's "alpinotype orogeny" (1924) — a term which was never intended to set up the Alps, with their extreme degree of nappe structure, as a standard.

and granitization. Indeed, it may be said of first order geosynclines that the beautiful sections of folded and overthrust fossiliferous sediments which figure so prominently in our concepts of geosynclinal mountains, are only a superficial skin and marginal zones, compared with the deeper and central portions of the geosynclinal body where compressive structure is coupled with even more fundamental transformation brought about by metamorphism and granitization. Folded geosynclines which lack this profound transformation, are of second order (miogeosynclines[18]), and it is more than coincidence that nearly all of them also lack another sign of mobility of highest degree, namely the welling-up of sima-born basic magma during their time of subsidence and sedimentation.

Where in the inner portions of eugeosynclines the mechanical forces of orogeny act most strongly and the rocks suffer intense differential deformation, *dynamic regional metamorphism* results. However, if the analysis in Part I, chapter IV, is correct, this process makes *only lower-grade crystalline schists* (cf. definition of dynamic metamorphism in footnote no. 2, p. 681) and, except at very great depths, *higher-grade schists* form only where *additional heat* is supplied by the *agent of granitization.* This agent *combines chemical with thermal action.* Both are required to granitize, whereas the latter, where acting alone, results in *higher-grade regional metamorphism,* which thus is a *byproduct of granitization.* The most fundamental transformations which occur during orogeny in eugeosynclines, are thus due to the chemical and thermal action of the granitizing agent.

Mechanical deformation and chemical-thermal action present themselves as different aspects of one major geological phenomenon, i.e. eugeosynclinal orogeny, and a close genetic relationship must exist between them. For not only is granitization confined to eugeosynclinal zones (as can be demonstrated at least for post-Archaean time), but also it is entirely absent during the pre-orogenic period of geosynclinal subsidence when sima-born basic volcanism has the stage. Obviously, granitization is either a direct consequence of orogeny, or it is itself an integral process at the root of orogeny. Like granitization, regional metamorphism of low as well as of high grade does not seem to occur in pre-orogenic time. The hypothesis that during the time of subsidence "geosynclinal metamorphism" or "load metamorphism" can be produced statically by load and geothermal increase of temperature, can hardly be reconciled with the numerous sections of completely non-metamorphic *geosynclinal rocks five and ten miles thick.*

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[18] Stille (1940) has divided orthogeosynclines into "eugeosynclines" and "miogeosynclines" and Kay (1944) has adopted a similar classification. Eugeosynclines have basic volcanism during their period of subsidence ("initial volcanism" of Stille, 1940), miogeosynclines do not. This criterion in most cases coincides with that of presence or lack of metamorphism and granitization.

Static granitization occurs in the same environment as synkinematic granitization, i.e. in eugeosynclinal belts. It does not occur unless strong orogeny has previously taken place. It thus appears to be *genetically linked with orogeny,* although it is later and therefore static. Its connection with a previous period of orogeny is further emphasized by the fact that it has commonly been preceded by synkinematic granitization in the same general area. In fact, in numerous regions static granitization is merely *the continuation, after the end of orogeny, of a process begun during orogeny.* A continuous evolution of structural forms of metasomatic replacement is characteristic. Typical cases have been described in chapters II and III, and many others could be added. In some other areas, there is a certain discontinuity between synkinematic and static granitization, indicating postorogenic subsiding and subsequent new rise of the front of granitization (cf. chapter Ia).

There are also *regions in which only static granitization is exposed.* In Part II, a case has been described where synkinematic granitization is absent not only in the immediate area of static granitization, but where in a much wider region synorogenic alteration has not gone beyond low-grade (dynamic) metamorphism. The question is whether in such cases there really has been no synkinematic granitization at all prior to static granitization, or whether it is merely concealed at depth. The second alternative seems to have much in its favor. As has been stated above, static granitization has frequently been preceded in the same area by synorogenic granitization, and always by orogeny. It therefore appears that orogeny and the birth of granitization are genetically linked. Moreover, static granitization has frequently spread beyond the limits reached by synkinematic granitization in the same area, and it is plausible that it has done so not only laterally but, to an even larger extent, vertically. For after the end of orogeny, mechanical conditions favor a *further rise of granitizing emanations.* The compressive stresses of orogeny have then disappeared, and general uplift in the geosynclinal belt tends somewhat to loosen the cohesion of the rocks.

That in postorogenic time the granitizing agent may actually penetrate higher levels than during orogeny, is proven by *cases of static granitization "at unusually shallow depth"* (cf. Part II). The contrast between more deep-seated and shallow static granitization is to a large extent responsible for the textural variability of the products of static granitization which, as a whole, is greater than in synkinematically granitized rocks all of which converge toward granitic gneiss,[19] whereas some products of shallow static granitization never reach the eugranitic-appearing stage even if their chemical alteration has been very thorough. Where static granitization can stratigraphically (and texturally) be demonstrated to have occurred at a very shallow depth, no synkinematically granitized rocks are visible. This is hardly a coincidence. In such cases, the synkinematic gneisses are probably concealed at greater depth. Where synkinematic and static granitization are exposed in the same area, the latter does not belong to the variety characteristic of very shallow depth. Where static granitization does not belong to that variety either, and where yet no synkinematic gneisses are exposed, they may be supposed to be present in the deeper portions of the geosynclinal belt although there is, of course, no proof.

The sizes of synkinematically granitized bodies known to me are all fairly large. At least, I have never encountered small and isolated patches of synkinematic gneisses. The process acts on a regional scale, and frequently reaches the dimensions of large batholiths (cf. Part I). *The sizes of individual statically granitized bodies* vary greatly (cf. Part III, chapters I to III). Very small bodies are common, though there are usually a number of them scattered in the same general area. Larger bodies are equally common, and frequently reach the dimensions of large batholiths. I think that *the majority of the large granodioritic and granitic batholiths have most*

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[19] Contemporaneous differential deformation is another factor which tends to make the fabrics of synkinematic gneisses more uniform, being unfavorable to the preservation of pre-existing grain textures.

probably formed by metasomatic granitization. Apart from the growing body of regional evidence for granitization, which the workers in this field are contributing in different parts of the world, certain general considerations seem to favor a metasomatic as against an igneous intrusive origin. Such arguments have been advanced by several authors; I refer to the recent reviews by H. H. Read (1943, 1944, 1948). These arguments include the room problem which is linked with the usual lack of signs of displacement of the country rocks near large batholiths; and the great predominance of granitic over basic rocks within the crust, in contrast to the even greater predominance of basalt and related rocks among the volcanic rocks—an argument which is strengthened by definite evidence of simatic origin found in some basalts. I would add another argument. It seems difficult to conceive of enormous masses of granitic igneous magma rising to shallow levels in the crust, in broad fronts extending over tens of thousands of square miles, without breaking through to the surface. If the large granitic batholiths were igneous, I should expect that those geosynclinal belts in which they occur, would in postorogenic time have been drowned in floods of rhyolites, dacites, and related rocks.

True igneous granitic bodies where an intrusive mechanism can be demonstrated and a genuine igneous texture is developed, are usually comparatively small. They may occur far away from areas of granitization and are, in fact, not restricted to eugeosynclinal belts. In other cases, *intrusive granitic bodies* occur *in areas of granitization* where they are often restricted to small-scale features such as dikes (cf. Part III, chapter III). Such cases may be interpreted as an advanced stage of granitization during which metasomatized material became finally mobilized and was thus able to intrude. This idea has been suggested by a number of authors. J. J. Sederholm (1926, p. 135) wrote of "anatexis" and "palingenetic (i.e. re-born) magma," implying essentially a process of re-fusion. C. E. Wegman (1930; 1935, pp. 334-335) uses the term "diapirite" for large masses of migmatitic material which plastically flowed and became intrusive. M. Reinhard (1935) speaks of "migma," i.e. mobilized migmatitic material which consists of a mixture of solid and liquid and might intrude like magma. H. G. Backlund (1937, p. 234) uses the term "rheomorphism" for partial or complete liquification of metasomatized rocks which then might act like magma. G. E. Goodspeed (1939, p. 399) speaks of "neomagma," i.e. locally newly formed magmatic material of metamorphic origin which performs a mass movement of plastic flow. He (1948, pp. 72-73; 1949) defines "rheomorphism" as mass movement of material of metamorphic derivation, and presents evidence in the form of "rheomorphic dikes."

The principle of mobilization and intrusive movement of material of metamorphic derivation does not necessarily imply whether the material is a plastic crystalline rock, a crystal mesh with subordinate interstitial liquid, a melt-solution, or a mixture of the two latter. Probably these cases all occur. That plastic crystalline rock is capable of local intrusive movement, even without having necessarily been granitized, has been mentioned in chapter III (p. 690). So far, rheomorphism has been definitely demonstrated only on a small scale (Goodspeed, 1949). It seems probable that generally intrusive bodies of rheomorphic origin are relatively small, compared with the size of large batholithic masses formed by metasomatism in situ. However, much field work still remains to be done.

As stated above, granitization comprises both *chemical and thermal action.* The latter involves introduction of heat conveyed by the mobile agent which infiltrates and granitizes the geosynclinal rocks (cf. Parts I and II). Except at very great depths, introduction of heat is required for regional granitization as well as for regional recrystallization in high-grade mineral facies. Probably additional heat is generated in situ by exothermic reactions (cf. V. M. Goldschmidt, 1921, p. 135; J. A. W. Bugge, 1945, p. 15), *after* higher temperatures have been attained. The *primary* source of heat is obviously from below. "From below" implies from depths greater than are exposed anywhere, i.e. exceeding 15 to 20 miles. Moreover, the source of heat must be at least as deep as the place of origin of the critical elements carried upward by the granitizing agent. I shall try to show that the added alkali originates not only beneath the geosynclinal prism, but most probably even beneath the lower boundary of the sial zone. The primary source of the heat required to granitize the inner zones of eugeosynclines, thus appears to be at very great depth. The time when this heat is generated beneath geosynclines and rises into them, is that of orogeny; for orogeny and birth of granitization are contemporaneous and genetically linked (cf. above). The question whether orogeny is responsible for generation of the heat, or whether both are consequences of something else, or whether generation of heat is a primary cause, leads into the realm of speculation. Perhaps geophysical, and especially seismic, methods may help to solve this question.

Chemical action involved in granitization includes both an *addition* of certain substances to the geosynclinal body, and *redistribution* of some of the material contained in it. We may speak of a "metabolism of geosynclines." Redistribution implies mobilization, transfer, and re-precipitation. Various elements may be so affected (including rarer ones). They will frequently, though not necessarily, be excess substances in rocks undergoing granitization ("excess" relative to a granitic composition).

Next to the alkalies, *silica* is the substance most frequently precipitated during granitization. Though part of the sediments in geosynclines have sufficient or excess silica, the granitization of many geosynclinal sediments and nearly all geosynclinal volcanic rocks requires additional silica, often in considerable amounts (cf. Part II, chapter IVe). A large amount of silica will usually be provided by transfer from silica-excess sediments, but it appears very likely that additional silica is introduced from beneath the geosynclinal prism. This is suggested by the fact that the granitizing agent has in most cases been able to satisfy the needs of silica-deficient rocks, irrespective of the composition of the geosynclinal column in individual areas. Also, in some regions quartzitic members have escaped granitization and thus kept their excess silica. Granitization of siliceous rocks obviously requires mobilization and transfer of some of the abundant aluminum of geosynclines.

The most important material added is *alkali* (cf. Parts I and II). The considerable amounts of alkali consumed in regional granitization could not be supplied by mere redistribution of the alkali originally contained in the geosynclinal rocks. For as a whole the geosynclinal column is alkali-deficient in comparison to granitic rocks, and there is no evidence of a regional loss of alkali in the non-granitized portions of geosynclinal bodies. Moreover, among sediments alkali is chiefly contained in argillaceous rocks, in which potash exceeds sodium. Therefore, even if the total alkali-deficiency is ignored, the contrast between regional sodium and potash metasomatism could not be accounted for. The only sediments very rich in alkali are salt deposits. Could they have supplied the alkali required for granitization? Theoretically, this might be possible in exceptional cases though no supporting field data are known. As a general solution, this hypothesis fails. There are not enough salt deposits, and in geosynclines they are less common than in non-geosynclinal basins, and in a number of cases entirely absent. Moreover, cases of strongly potash-predominant metasomatism could not be explained, in view of the scarcity of potash salts in salt deposits.

It follows that most of the alkali which is added during granitization, has come *from beneath the geosynclinal prism.* Could it have been derived from the gneissic and granitic pre-geosynclinal basement rocks? This appears unlikely. There would be required a process of "de-granitization" of the basement, for it would have to give off a considerable part of its alkalies (and probably some of its silica) in order to make into granitic rock the overlying geosynclinal prism deficient in alkali. Melting of the basement would not solve the difficulty, since it would merely result in mobilization and rise of granitic material, but could not create an agent sufficiently enriched in alkali to perform regional granitization of alkali-deficient rock masses. As to observed facts—quite apart from the question whether *all* eugeosynclines do have old granitic basements *throughout* their inner portions—the granitic basement has remained granitic wherever it is actually visible in eugeosynclines, which is partly in their inner and more often in their outer zones.

It must be concluded that the alkali added to the geosynclinal prism has come *from non-exposed depths,* and probably from beneath the granitic or sial zone. P. Eskola (1932; 1933) suggests that granitic magma may form by *selective* re-fusion of rocks in the sima zone, through formation of a granitic pore magma which is squeezed out and rises into the geosynclinal belts. This revolutionary idea of a selective mobilization might perhaps be somewhat modified, by assuming a comparable process, which, instead of selective remelting, might consist of selective mobilization of ions and molecules and thus produce an alkali-rich mobile phase, some kind of emanation which by means of infiltration, permeation, and diffusion would rise into the geosynclinal prism and perform metasomatic granitization. Granitic magma rising into the geosyncline could intrude, inject, and locally metasomatize, but it would not be chemically capable of supplying the large amounts of alkali which are consumed in large-scale regional metasomatic granitization. C. E. Wegmann (1935, pp. 346-347) suggests that deep-seated melts considered to be the source of the substances added during granitization, might be of an alkaline type and, when intruding geosynclinal rocks, give off alkali, thus cause migmatization and themselves become calc-alkaline. This idea might be modified by assuming that rocks of an alkali-basaltic composition are common in the upper part of the sima zone, and that such rocks would, as a result of selective mobilization, give off alkali-rich emanations. Tempting though this idea is, there appears to be a major difficulty. The basic volcanic rocks of deep origin which are so characteristic of the pre-orogenic period of eugeosynclines, are mostly *not* alkaline, nor can they have been and lost alkali to the geosynclinal sediments, since at the time of basic geosynclinal volcanism no granitization and no regional alkali-metasomatism are known to occur.

Whatever ideas may develop as to the origin of the agent of granitization, a few things seem to emerge from the realm of speculation into that of fact, namely *that the majority of large granitic batholiths appear to be of metasomatic origin; that no magmatic source of the agent of large-scale regional granitization is visible within the depths exposed; that the birth of granitization is genetically connected with eugeosynclinal orogeny; that its primary motor is at great depth.* In this sense, *synkinematic granitization is the original process and static granitization a more secondary consequence, although it is able to rise to higher levels than are generally reached by synkinematic granitization.*

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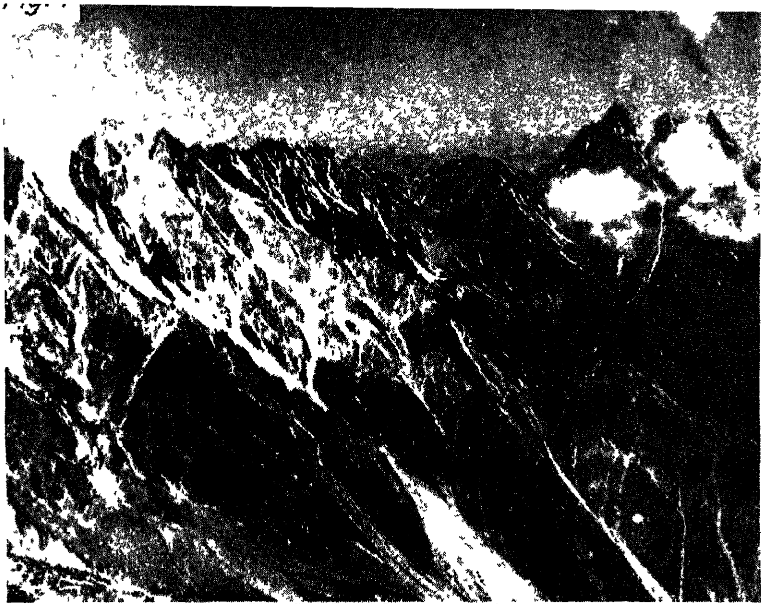

Fig 2

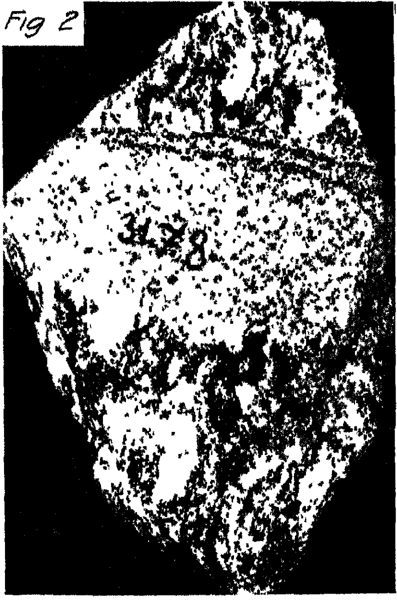

Fig 3

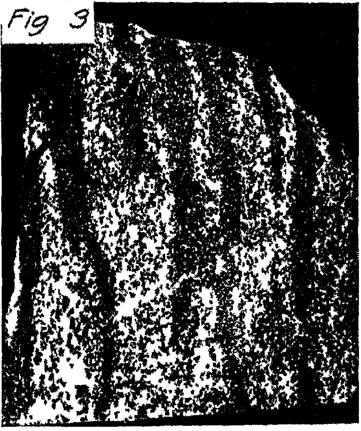

PLATE 1

Fig 1 Postkinematic white tourmaline-granite in synkinematic darker granitic biotite-gneiss East face of Gunar Peak (17,520′) northwest of Nanga Parbat, Northwest Himalayas The face is about 5000′ high After Misch (1935), Bild 42

Fig 2 Small replacement dike of white tourmaline-granite in synkinematic granitic biotite-gneiss Lower border is irregularly sutured, upper border is fracture-controlled and sharp East side of Buldar Ridge north-northeast of Nanga Parbat (no 31 7.8) About natural size

Fig 3. Replacement hornblende-quartz-diorite in weakly schistose hornfelsic amphibolite Metanorite mass northwest of Nanga Parbat gneiss massif North of Gor Pass, north of Indus R. (no. 29 5 9). About ⅔ natural size.

ORIGIN OF PIMPLE MOUNDS

E. L. KRINITZSKY

ABSTRACT. Pimple mounds are remains of dunes and hillocks which have formed under a wide variety of conditions. These include environments of glacial outwash, coastal sand flats, and sandy point bars of meandering streams.

I. INTRODUCTION

PIMPLE mounds of various areas, and differing names, exist in great abundance. Along the Gulf Coast of eastern Texas and southwestern Louisiana they are best known and are numbered by the hundreds of thousands. Also they are plentiful in southeastern Missouri, western Washington, and parts of Iowa, to mention but a few regional areas.

Much has been written on the subject of these interesting features, but authors in the past have limited their studies to restricted areas. Thus there have been developed theories of origin which are based on purely local evidence and which become inadequate when applied to identical forms elsewhere. Without doubt, pimples may have a variety of possible origins only a few of which are of widespread importance.

The writer hopes to demonstrate a principle of pimple mound origin which has wide application but which, up to now, has not been advocated.

II. ORIGIN OF PIMPLE MOUNDS

A. DESCRIPTION

Typical pimples generally have the appearance of low, flattened domes which are either rudely circular or elliptical in horizontal cross-section. Often they coalesce and are ridge-like in appearance. Their basal diameters vary between ten and over a hundred feet and their height ranges from less than one foot to over five feet. In general, pimples of the Gulf Coastal Plain average between 35 and 50 feet in diameter and two or three feet in height. In all areas where these mounds occur, they are in extreme abundance and occupy large percentages of the land surface. Isolated individual mounds are very seldom found.

In the Gulf Coastal region, the material of which these mounds are composed is commonly a fine to sandy loam which is coarser than, and entirely distinct from, surrounding soils.

In Washington (Dalquest and Scheffer, 1942) the mounds are of a black graveliferous silt situated on a stratified, yellowish graveliferous alluvium deposited by glacial outwash. In portions of Missouri and Arkansas the mounds are composed of coarse loam in areas of distinctly finer soils.

Fossils or artifacts are hardly ever found in these hillocks.

B. MEANDER BELT PIMPLE MOUNDS

In plate 1 there is shown the appearance of typical pimples situated on Recent alluvial deposits in the Advance lowland near Cape Girardeau, Mo. It can be seen from the still distinct river scars that this entire surface has been reworked by a meandering stream which has since migrated away from the area. Typical meander bends, swales, and accretion topography on point bars are easily distinguished. And, associated with these features, are pimples oriented in patterns which are entirely conformable to the accretion ridges developed on point bars of the meandering stream.

The above conformity of pimples to meander belt topography is by no means unique. Similar situations are found in many areas. Typical examples occur in the Red River valley in Oklahoma and Texas, and in flood plain areas of Arkansas and Tennessee.

Accretion ridges on growing point bars along much of the lower Mississippi River are typically sandy, as are many towheads. These are sculptured by river currents during high water, built up as a result of accretion during falling water, and left exposed when lower water levels are reached. Since these accretions are periodic and occur generally after intervals during which the river bends have been migrating, most sandy accretion ridges are separate and distinct. Between these accretion ridges are the so-called swales which develop luxuriant growths of willows and trap fine sediments when flooded during times of high water. Thus the topography contains very gentle swells of coarse sediments protruding from areas filled with finer alluvium. In very short order, mantles of vegetation creep onto these sediments and develop soil profiles into them. To a great extent, vegetation aids in preserving certain of the topographic configurations. With removal of this covering by agriculture, many meander belt features, such as are exhibited by soil fertility and topographic variations, come into prominence again.

Plate 2, fig. 1 presents the appearance of a sandy towhead, as well as sandy accretion areas, on a point bar south of Thomastown, Louisiana. All of these features have been removed from the area of active river action. Eventually vegetation will develop over and preserve these sandy surfaces as it has done already elsewhere in the finer sediments. It can be noted from the photograph that both the towhead and the sandy accretion strip contain distinct sand mounds of fairly uniform size. These have been formed chiefly by eddies and turbulent flows resulting from river action during time of flood and perhaps subsequently modified slightly by wind. It can be noted that the individual magnitude of these sand mounds is the same as is that of typical pimples elsewhere. Likewise, the orientation of these mounds along the pattern of a meander curve is the same as are the arcuate rows of pimples shown in plate 1.

Plate 2, fig. 2, depicts a similar situation in which there has developed a silty sand flat in an abandoned course of the Mississippi River. The locality is near Alligator Lake in Louisiana. It can be noticed that here too there are sand mounds having the individual magnitude of pimples. Many have elliptical shapes and have coalesced. It is observable here that orientations of these mounds are in two directions roughly at right angles. When such dual orientation occurs on limited portions of a fairly extensive sand flat and, if at the same time the mounds are equally spaced geometrically, it is apparent that they may appear to have no particular orientation at all—once they become subdued and incorporated into a more blended topography. This is illustrated in the case of pimples in plate 3.

Plate 3 is an aerial view of the surface at Vincent Island in Louisiana. The island is Prairie Terrace, Pleistocene of age, and therefore independent of the Recent marsh conditions which can be seen to prevail in the northern portion of the view. During the late Pleistocene, this terrace surface carried a silt laden, meandering stream which developed many silty flats such as the channel filling pictured in plate 2, fig. 2. In the south central portion of plate 3 meander scars are still visible, and many coalesced pimples can be seen to have the dual right angle orientations identical to those of the sand dunes of plate 2, fig. 2.

Thus it becomes apparent that meandering behavior of rivers tend to develop silty and sandy concentrations of sediment in individual strips which eventually cover much of the total surface area within a meander belt. As these coarse deposits are forming, their surfaces are sculptured with water-carved mounds. Vegetation quickly moves in to form a protecting cover and, with the formation of a soil profile, these mounds become our familiar pimples.

C. COASTAL PIMPLES

Since the pimples of meander belts have originated from sand mounds on bars, and so forth, it is logical to expect pimples to have been derived also from eolian sand dunes on coastal beaches. The latter is found to be the case

As a typical example of this sort, plate 4 shows the arrangement of pimples in southern Chambers County, Texas. It can be seen from the inset photograph that this is an area of a former sand spit. Outlines of several stages may be observed from the patterns of ancient beaches at least three of which are yet distinctive. During the times at which these beaches existed, the spit was much longer than at present, its southern part having been since truncated. At the present time, the seaward part of this peninsula contains marshes which present quite a different environment from that which formerly existed. Hence an environment for the formation of dunes is no longer present. But, it can be seen that, during earlier times, there were sandy beaches in this area and upon these beaches, small, eolian sand dunes were formed and preserved. Since these beaches were in a process of migration, their dunes have become spread over a large area but still maintain their orientations parallel to the former shorelines. And with moderate changes occurring over a period of time, these coastal sand dunes have become typical pimples.

The uniformity in size of these former dunes may be related to uniformity of beach areas upon which they have originated. Only within widespread deserts and coastal beaches containing abundant sand deposits are very large dunes able to develop. With restrictions as to area of sand flat and supply of sand, dunes tend to be small in size. An environment of this sort explains the size homogeneity of pimples in this area. Likewise, more abundant sand deposits along beaches in other areas, which would thus develop few, but very large, dunes, explain the absence of pimple mounds in many sand areas elsewhere.

D. MOUNDS IN GLACIAL DEPOSITS

In certain areas of glacial outwash in western Washington there are the well known "Mima mounds." With regard to size, shape, orientation, and relationship to alluvial deposits, these are essentially similar to the pimples of the Gulf Coastal Plain (cf. Dalquest and Scheffer, 1942).

These mima mounds are formed of coarser materials than are the Gulf Coast's pimples, however their association with glacial outwash channels and their situation on stratified outwash gravels speak for a similar fluvial origin. There is nothing in their physical characteristics to set them apart as unique.

Likewise in other glaciated areas there are features which very closely resemble ancient mound forms. Gwynne (1942, p. 204) has published an airphotograph of an area southeast of Ames, Iowa, which very distinctly contains a pimple pattern although the appearance is much subdued. In this area there are swell and swale features which are interpreted as resulting from oscillations of an ice front of the Mankato lobe of the Wisconsin drift plain.

III. FORMER HYPOTHESES OF ORIGIN

Suggestions as to the origin of pimple mounds have been put forth by a number of authors. Following are some of the theories together with certain objections as to their universal validity.

1. Legends are the most interesting of the available hypothesis and are quite as logical as are some views expressed by cultured persons. Following are two folk stories reported by Harris and Veatch (1899) regarding pimple mounds of Arkansas, Louisiana, and Texas:

> "An old darkey offered the following explanation: when the 'Great Massa' made this earth he made it with a sieve and when he finished the sifting there were a number of lumps left in the sieve. These were thrown out forming the mounds. Another explanation . . (was) . . given by the Caddo Indians. Their legend ran that many, many years ago the country was inhabited by a race of giants. For some reason, which the legend did not state, the giants were

carrying dirt in their aprons; when the dinner horn sounded they dropped the dirt where they stood and hurried away to a dinner from which they never returned."

Objections to these views will be at once apparent to many people.

2. Many brief and generally unsupported views have been presented by writers during the latter half of the nineteenth century. Following are a few of these:

a. According to Marquis de Nadaillac (1895), pimple mounds are of human origin being either garden beds, tepee or wigwam sites, or burial mounds. It is difficult to see how such a dense population as would have been required could ever have lived in these areas and the absence of artifacts or human remains excludes the likelihood of human influences.

b. Hilgard (1873) attributed the mounds to a species of ant which has since disappeared. And Veatch (1905) extended this view to include termites. However, the structures within these mounds show no evidences to this effect and the presence of gravels within and upon pimple mounds of Arkansas and Washington preclude the possibility.

c. A purely wind origin for pimple mounds was recommended by Featherman (1872) and Clendenin (1896). However their hypotheses require an arid climate and in no way do they explain meander patterns, uniformity of mound size, and so forth. Also their views regarding masses of earth lifted by the uprooting of trees in storms and modifications by burrowing animals do not explain greater coarseness of material and the geographic limitations to areas of certain alluvial deposits.

d. Various forms of an "eruption" theory have been suggested by several authors including Hopkins (1870), Clendenin (1896), and Harris and Veatch (1899). These involve springs or aqueous volcanoes resulting from artesian pressures, origins through the action of gas vents, and bulges due to the unequal weight of uneven clay layers on plastic, water-logged sand lenses. Likewise, these modes of development do not explain uniformity of sizes, meander and ancient beach pattern arrangements, etc.

e. Brief statements have been made to the effect that pimple mounds have been produced by a rapid rush of waters and secondly by a slow differential erosion (Harris and Veatch, 1899). Neither of these views are sustained by facts.

f. Branner and Purdue (cf. Veatch, 1905) have suggested that pimple mounds represent immense concretionary formations and others have advocated unequal weathering of limestone, and even fish nests exposed by elevation. However, these views are too improbable to merit consideration.

3. The mima mounds of Washington have been discussed in an interesting paper by Dalquest and Scheffer (1942). These authors effectively refute earlier theories of origin which have included views that debris was washed into pits on still or moving ice, and that silt, when frozen, cracks into huge polygons which can absorb sediments, and so forth. However their recommendation that these mounds are purely the work of gophers heaping up earth over many generations is equally untenable. Were pimple mounds formed in this manner, it is not likely that they would be limited to areas of alluvial deposits. A shallow lying base of gravels, into which gophers will not burrow, is an integral requirement of the gopher-hypothesis. And the authors suggest that in parts of the country other than Washington a clay hardpan or bedrock may substitute for dense gravel. If this hypothesis is true, then pimple mounds should be common features on thin residual soils overlying bedrock wherever there are gophers. This definitely is not the case. To the writer's knowledge, the most typical pimples are restricted to alluvial deposits and nearly always reflect configurations of these deposits. Gophers cannot account for these essential features.

Also, a strong argument, put forth by Dalquest and Scheffer in favor of gophers, is that "mima mounds" are found on terrace slopes and that other modes of origin cannot account for such occurrences. However, an alluvial origin can be shown to be the case in areas where excessively steep river banks may have the appearance of terraces. This is especially the case when banks are abandoned through river shiftings and become subdued with vegetation. Plate 5 contains an example of distinct sand mounds which were formed by eddying currents during a flood of the Mississippi River and have been left plastered more than midway up a bank over 35 feet in height. It is not improbable that similar features may have formed along swollen glacial streams in western Washington.

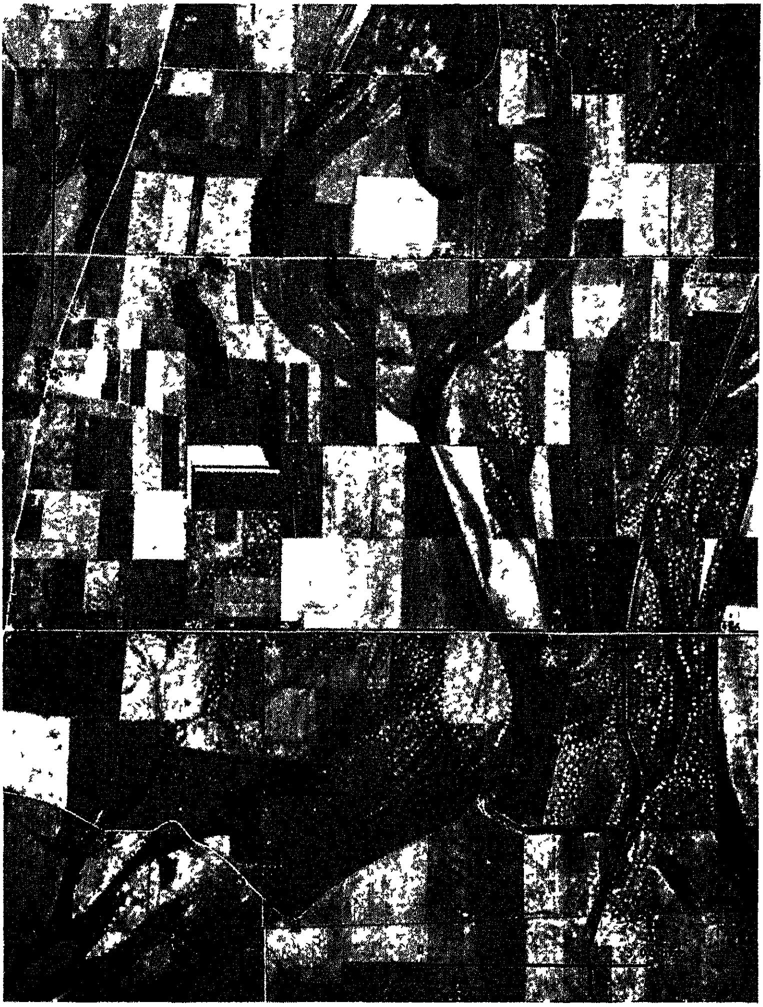

PLATE 1 Pimple mounds at north end of Advance Lowland, near Cape Girardeau, Missouri Note mound alignments in the pattern of sand accretion ridges on former point bars of a meandering river.

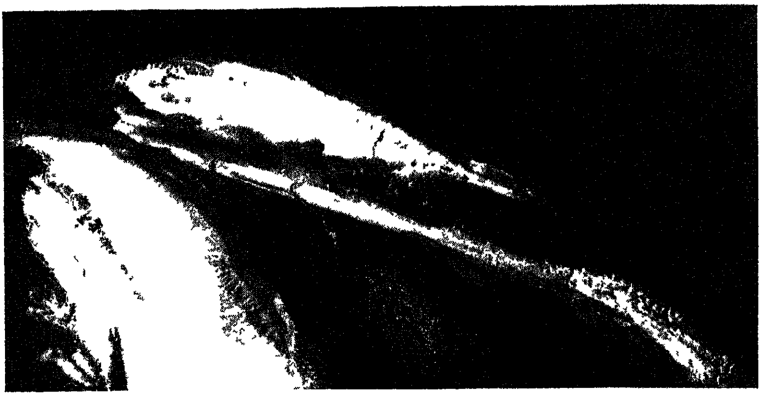

PLATE 2, Fig 1 Towhead and sand deposits in a partly abandoned river channel Mounds on the sand deposits have similar shapes and alignments as have pimples S of Thomastown, La.

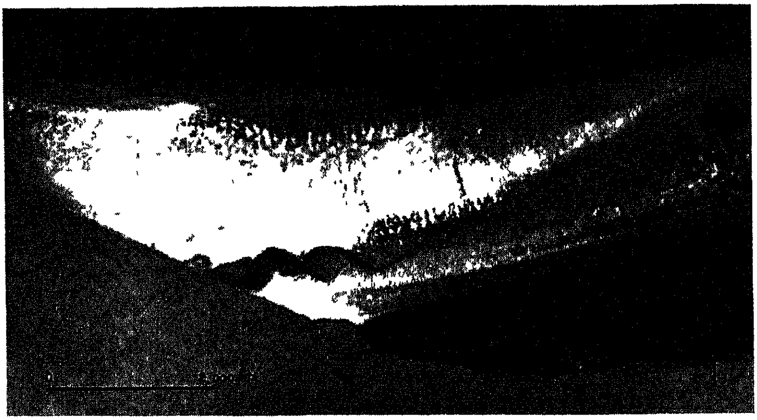

PLATE 2, Fig. 2. Sand mounds formed in an abandoned river channel. Note magnitude and alignment in two directions 4 mi east of S. tip of Alligator Lake, La

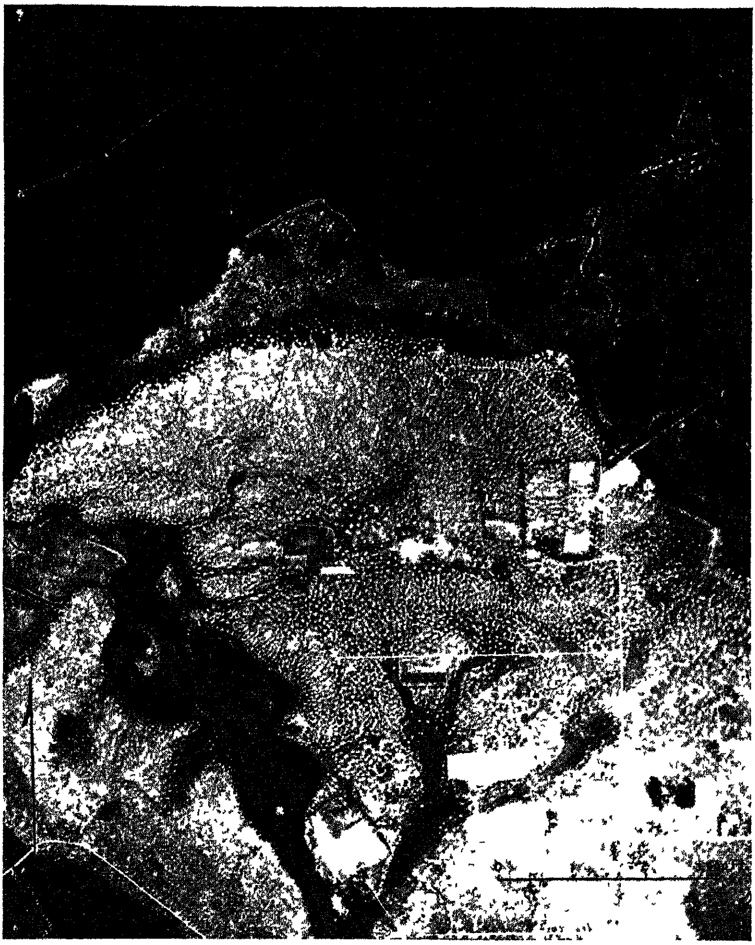

PLATE 3 Pimples on the surface of Vincent Island, Louisiana, surrounded by marsh to the north and east Northward is Black Lake Pimples pass under the Recent marsh deposits

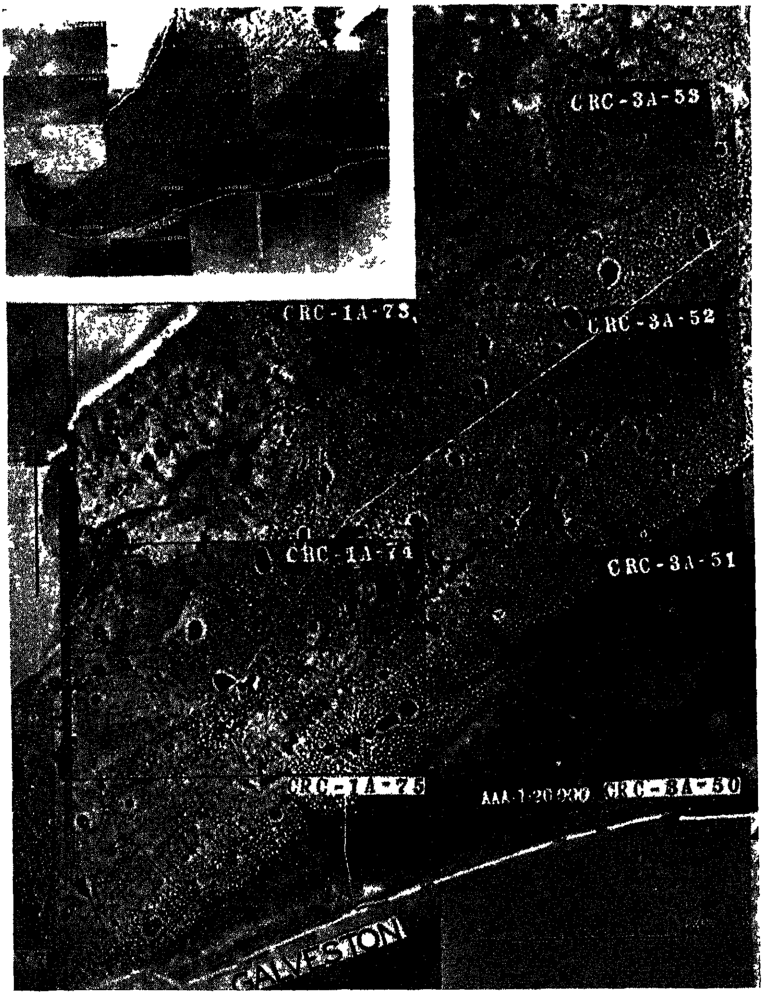

PLATE 4 Pimple mounds in southern Chambers County, Texas Note alignments along old beach ridges and the previous forms of sand spits. AAA photo.

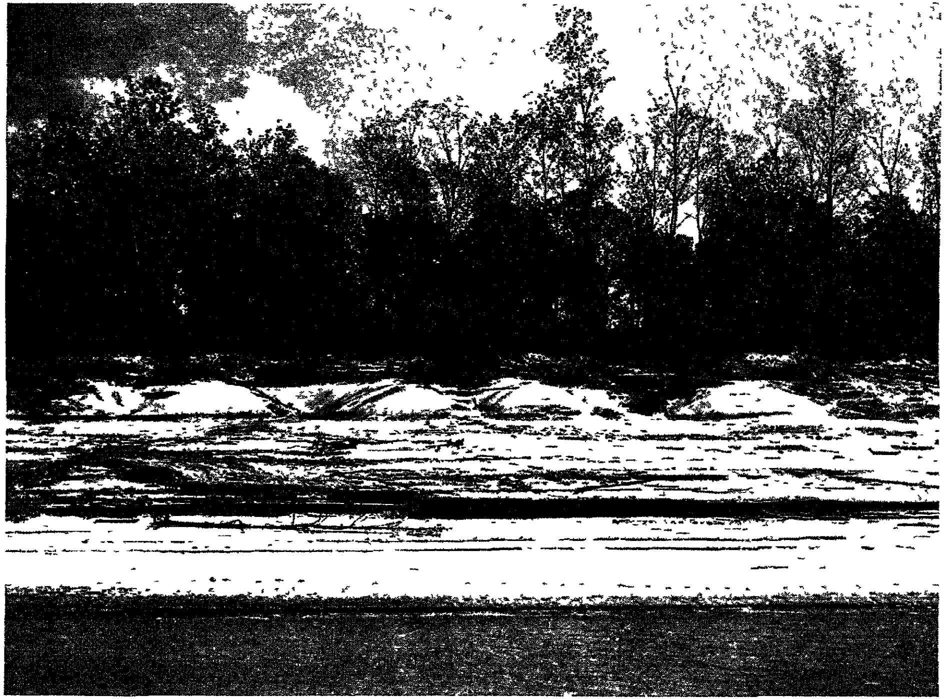

PLATE 5 Sand mounds formed by eddying currents during high water stage. These are plastered along a 35-foot bank slope East Bank, Mississippi River, Avondale Bend, Jefferson Ph., Louisiana, 1944

4. A recent theory, which has the merit of being still different from the flood of views already mentioned, is one proposed by Dietz (1945). His suggestion is that the mounds "are water-deposited features originating as sandy islets along the margin of marsh-grass-lined lagoonal swamps. Aerial observation . . . shows these islets along the shore in various stages of formation and grading into mounds preserved on land." This condition is illustrated in plate 3 where it can be seen that pimple mounds extend out into the marsh bordering Black Lake. In the left portion of the view, pimples can be seen to protrude from out of the water. However, this situation does not illustrate growth stages. The pimple mounds in this area are Pleistocene in age and have in part been alluvially drowned by an encroachment of Recent marsh deposits. On the periphery of this encroachment, they protrude from under the Recent covering. Thus, this gradational relationship has been misinterpreted in explaining stages of origin. That Recent deposits have flooded onto Pleistocene surfaces has been proven in bore holes by the presence of oxidized materials directly below the black carbonaceous Recent marsh materials. And, finally, an origin of pimple mounds by development in lagoonal marshlands cannot possibly explain the Missouri pimples in plate 1, or those of western Washington.

The writer does not deny that these processes suggested by earlier workers are adequate for producing pimple mounds. However, their application is restricted and is of the much lesser importance when compared with the much more widespread occurrence of river mound and eolian dune pimples.

IV. SUMMARY

1. Pimple mounds reported from various areas, such as Washington, Missouri, Arkansas, Texas, and Louisiana, are essentially similar.
2. In meander belt areas, pimples conform to patterns of accretion ridges on point bars.
3. Sandy towheads and point bars are observed as containing water and wind developed sand mounds which have identical sizes and orientations as have pimples.
4. It is suggested that such mounds are preserved by mantles of vegetation and later are exposed as the familiar pimple features.

5. In coastal sand-flat areas, dunes are observed to have orientations which indicate the beach origin of typical pimple mounds.

6. Glacial outwash areas, such as in Washington and Iowa, have developed mound forms on their surface which are identical to typical pimples.

7. Pimple mounds, in conformity to the eolian dune and river mound hypotheses of origin, are generally restricted to areas of alluvial deposits.

8. These hypotheses of origin appear to have more universal applications than do other modes of pimple mound origin previously advocated, although the latter theories may remain valid in many special instances.

V. CONCLUSION

Important modes of origin of pimple mounds are processes responsible for sand dunes along coastal beaches, sand mounds on accretion topography formed by meandering streams, and alluvial mounds deposited from glacial outwash.

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CERAMIC PRODUCTS; THEIR GEOLOGICAL INTEREST AND ANALOGIES

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ABSTRACT. Some phenomena of the ceramic industries are analogous to processes believed to operate in igneous, metamorphic, and sedimentary petrogenesis. The shaping of clays and the deformation of rocks produce comparable petrofabrics. The interaction of igneous and country rock is paralleled by that of silicate-melt and refractory. Fronts and culminations exist in slagged refractories, but there is no evidence that they are produced by solid diffusion. Ceramic phenomena that suggest the operation of solid diffusion are, however, recorded.

INTRODUCTION

THE ceramic industry depends largely on the exploitation of rocks and minerals; among its raw materials clay is pre-eminent. The literature devoted to clays is very extensive; those parts of it which are of general interest to geologists are on the whole, relatively accessible and well-known. There exists, however, further literature, scattered through a variety of journals, which is less familiar. Admittedly this literature varies much in scientific value, but in it are described phenomena arising during the manufacture and use of ceramic artefacts that show suggestive analogies, real or apparent, to geological processes. The books by Chesters (1946), Eitel (1941), Norton (1942), and H. Wilson (1927), the various reports on Refractory Materials issued by the Iron and Steel Institute of Great Britain, together with a recent review by Hilliard and McKee (1948) can be recommended to the geologist who wishes to extend his reading into the ceramic field; all these works have extensive bibliographies.

H. Wilson (1927) cites, in a classification, no less than 44 types of ceramic product. Much of the present communication, however, is restricted to consideration of monomineralic ceramic articles, English earthenware, and some important refractory compositions. Further, the interaction of refractories with slags, glasses, and gases is reviewed.

Bowen (1943) has already indicated points of contact between petrology and silicate technology, while Hilliard (1946) noted some analogies between geological phenomena on the one hand, and those encountered in the manufacture and use of refractories on the other.

ARTIFICIAL THERMAL METAMORPHISM

Tomkeieff (1940) has reviewed the nomenclature of the rocks that appear in thermal aureoles and has shown how a classification may be based either on composition or on texture. Benson (1945-46) has found it convenient to adopt that classification which carries a textural significance. Thus "buchite" is used for a vitreous or semi-vitreous material derived from argillaceous or arenaceous sediments, and "porcellanite" for the unvitrified, finely crystalline, metamorphic products of such sediments; "hornfels" is reserved for their less finely crystallised derivatives. The same classification can be conveniently adopted in the present text.

Many "fired" ceramic products, e.g., tableware or chemical porcelain, contain a quantity of glass with which are associated new or relict crysalline phases. They thus find their analogues in the buchites. It is significant that buchites tend to associate, as xenolithic or contact rocks, with minor intrusions, i.e., with intrusions that provided natural "firing" which was relatively short and thus, in duration, showed some measure of approximation to the "firing" of the potter's oven. Thermal treatment has been inadequate to ensure complete recrystallisation. Crystalline ceramic products resembling hornfelses do, however, exist. These frequently contain a single chemical species, with but few accompanying impurities, which has been heated to temperatures below its melting-point; the amount of low-melting material available for the formation of glass has been small. Such artefacts resemble the recrystallised quartzites and marbles; both the factitious and natural products owe their texture more to their monomineralic constitution than to the type of thermal treatment they have received.

The property of recrystallisation at temperatures below the melting-point is utilised in the production of certain ceramic articles, e.g., the well-known alumina ware; the history of this ware has been reviewed recently by Singer and Thurnauer (1947). Stott (1938) has noticed that MgO, BeO, and ThO_2 exhibit the same capacity for recrystallisation.

Recrystallised ceramic products are frequently described as "sintered." Sintering, however, appears to be distinct from recrystallisation, but may be assumed to be the precursor of the latter process and to merge into it as the temperature is raised. In the operation of sintering, a powder that has been compacted in the cold is heated to a temperature below its melting-point; the mass becomes rigid and coherent. Bangham (1947A; 1947B) has ascribed the coherence to the effect of forces, acting normally to the free surfaces of the particles, of the type which give rise to adsorption, adhesion, and friction. The rigidity of the solid allows these forces to operate initially only over a small area of two contiguous surfaces. This rigidity diminishes, however, as temperature rises; thus the increasing capacity of the solid to deform permits the forces to act over ever widening areas. Presumably the intimate contact finally produced allows the migration of ions from one particle to another and recrystallisation occurs. The natural recrystallisation of rocks would appear to imply an antecedent process of sintering.

Sintered ceramic ware is often shaped by casting and, in consequence, mechanical deformation of the grains does not occur. Buerger and Washken (1947), however, have studied the recrystallisation of strained solids, viz., CaF_2, MgO, $CaSO_4$, Al_2O_3, and BeO. They found that recrystallisation failed to occur below a critical temperature which, for a given material, decreased as the amount of initial deformation increased. The presence of a solvent lowered the temperature at which recrystallisation occurred. Such observations are of interest both to the ceramist and to the student of the dynamic factors involved in metamorphism. However, there seems to be scope for comparing them with the results of other workers who have devoted attention to the physical and chemical changes that can occur in the solid state (Hedvall, 1938A; 1938B). Thus Hedvall reports that the compressive strength of Fe_2O_3 is more markedly augmented by heating over the recrystallisation temperature if this substance has a faulty lattice initially (cf. the strained lattice produced in the compounds investigated by Buerger and Washken).

Of very direct interest to the geologist is the recrystallisation of quartz. This process could be expected to occur during the extended use of silica-bricks in the linings of industrial furnaces, but no record of it has been found in the literature. Dale (1924-25), however, produced a sandstone-like material, having marked cohesion, by heating ground quartz at ca. 1050°C. under a pressure of 270 lbs./sq. in. This observation suggests sintering.

D'Eustachio and Greenwald (1946) obtained clear evidence of the thermal recrystallisation of quartz under experimental conditions resembling those employed by Buerger and Washken (op. cit.). When quartz-individuals were stressed, they ceased to be crystallographically homogeneous but became polycrystalline associations of units having slightly different orientations. By appropriate thermal treatment the individuality was again restored. Certain structures observable in natural quartz, and presumably to be ascribed to the effects of stress, are significantly like the polycrystalline associations described by D'Eustachio and Greenwald.

CORDIERITE AND MULLITE

"Fired" clays and ceramic "bodies" may contain minerals that are to be found also in buchites and hornfelses. Most clays and bodies, however, are too poor in magnesia to yield cordierite, although this mineral is of frequent occurrence in the rocks cited. Cordierite may be formed, however, if the chemical composition of the ceramic material is adjusted by an appropriate addition, e.g., of talc. Thus Rooksby and Partridge (1940) ascribe the presence of cordierite in glass tank refractories to the use of magnesium compounds by the manufacturers. Shand (1943) found cordierite in a similar situation and believed that magnesium from the glass was involved in its formation. Hausner (1946) has lately reviewed our knowledge of cordierite-rich bodies. Phemister (1938-42) has described interesting occurrences of cordierite in the spent shale of oil-distillation retorts; in some instances this mineral appears to have been formed by simple thermal metamorphism.

A crystalline phase appearing in well-fired clay-containing ceramic articles is a compound that was long thought to be sillimanite. A similar compound had been encountered in buchites, e.g., in those associated with the Tertiary minor intrusions of Mull (Thomas, 1922). Thomas believed that the "sillimanitic" variety of these buchites had resulted from the simple fusion of a sediment having the composition of a highly aluminous fireclay. He directed attention to the comparable but artificial occurrences of "sillimanite" in glass furnace refractories as described, for example, by G. V. Wilson (1918). Bowen, Greig, and Zies (1924), however, were able to show that the "sillimanite" of both Thomas and Wilson was not $Al_2O_3.SiO_2$ but $3Al_2O_3.2SiO_2$; this compound was called mullite. At first there was some tendency to deny that mullite was a real compound, but Taylor (1932-33), among others, insisted that it was not merely an intimate mixture of sillimanite and corundum. Nahmias (1932-33) obtained evidence that there is no series of compounds between sillimanite and mullite, although the X-ray powder photographs of mullite showed some variation; these variations were attributed to the presence of TiO_2 and Fe_2O_3, rather than to differences in the Al_3O_3/SiO_2 ratio. A. J. Bradley and Roussin (1931-32), however, claimed that porcelains contained a material which they termed "porzite"; this, though similar to sillimanite and mullite, was distinct from them. Posnjak and Greig (1933) could not substantiate this hypothesis. The problem of mullite has been investigated more recently by Rooksby and Partridge (1939) who were able to distinguish three varieties of mullite by X-ray powder photography. Heating of alumina and silica in stoichiometric proportions yielded alpha-mullite $3Al_2O_3.2SiO_2$; compounds intermediate in composition between sillimanite and alpha-mullite appeared to be non-existent (cf. Nahmias, op. cit.). Alpha-mullite was not found among the natural mullites examined. When alumina was in excess beta-mullite was synthesised. The presence of iron and/or titanium in solid solution yielded gamma-mullite. Mullites from the classical occurrences in Mull were of the beta and gamma varieties. Most of the ceramic products examined contained mullite which varied in composition between the two extreme varieties, alpha and gamma mullite respectively, according to the amount of iron and titanium present in solid solution.

It is interesting to note that mullite is now prepared on a large scale by fusion-casting in the manufacture of, for instance, glass-tank furnace blocks. Rooksby and Partridge (op. cit.) have found beta-mullite in the cavities and blow-holes of such blocks. Hedvall (1948) has recently mentioned the synthesis of mullite by powder reaction in the presence of metallic oxide catalysts.

The crystallised alumina-silica compound, identifiable in ceramic products by means of the microscope, can be referred to mullite with considerable confidence. It is less certain that the "sillimanite" of metamorphic rocks, if the mineral has been recognised only on the basis of optical data, is not, in some instances, mullite.

It has now been shown that kyanite, andalusite, and sillimanite can be transformed into mullite by heating; the change occurs most readily in kyanite and least readily in sillimanite (Greig, 1926; Riddle and Peck, 1935). The relative stability of the three $Al_2O_3.SiO_2$ minerals can, indeed, be inferred from the mode of their natural occurrence. In the thermal metamorphism of pelitic rocks andalusite gives place to sillimanite as the metamorphism becomes more pronounced (Harker, 1939). Moreover, Williamson (1935), for instance, has described how the "regional" kyanite of a gneiss is supplanted by andalusite and sillimanite where the gneiss is contact-altered by a later intrusion.

The thermal transformation of the $Al_2O_3.SiO_2$ minerals into mullite has practical significance in the manufacture of refractories based on these materials. The change should have been effected by calcination before the refractory is used; otherwise failure under load may ensue at high temperatures. Failure depends on the change of the alumina-silica compounds to mullite; this change can occur at temperatures below those at which any liquid is formed in the system (Bowen, 1943).

METASTABILITY IN METAMORPHIC ROCKS AND CERAMIC ARTEFACTS

When the metamorphism of a mineral-association is produced by rising temperature and pressure there is a tendency towards the establishment of a new association of crystalline phases; this new association is called into being because the novel conditions demand a corresponding equilibrium assemblage of minerals. True equilibrium may not, however, be attained; thus, for example, andalusite and sillimanite may co-exist in hornfelses. When, on the other hand, temperature and pressure subsequently decline, phases that came into existence at the peak of metamorphism commonly persist; strictly, such phases are now in metastable equilibrium (cf. Harker, 1939).

The type of mineralogical reconstruction that is effected artificially in the ceramic industries can be equated most nearly with the process of thermal metamorphism. The mineral constituents of a ceramic artefact behave like those of a rock in their attempt to attain a new equilibrium in response to a rise in temperature. This attempt is patent, for example, in the relationship that exists between the porosities of fired articles of various compositions and the relevant equilibrium diagrams (Shelton, 1948). Mellor (1907) realised long ago that true equilibrium was often far from being attained—"The reaction between the different constituents of the body, in firing, is arrested before the system is in a state of equilibrium. The chemistry of pottery is therefore largely a chemistry of incomplete reactions . . . ." Mellor interpreted the word "reactions" very liberally in this context and used it to cover not only chemical but a variety of physical changes (Mellor, 1914C).

The presence of glass in a ceramic article is an obvious indication that a metastable condition exists. The persistence of tridymite and cristobalite in ceramic artefacts carries a similar significance; at room temperature quartz is the stable phase.

A PARTICULAR PRODUCTION-SEQUENCE AND ITS GEOLOGICAL ANALOGIES

The manufacture of British earthenware may be represented by the following much abbreviated flow-sheet:—

> Raw materials → mixing "ark" → sieves → magnets → "finished ark" → filter-press → pugmill → making (by shaping the plastic "body" or by casting in plaster moulds) → drying → biscuit-firing → glazing → glost firing.

The raw materials are Cornish china clay, "ball clay" from Tertiary deposits in Devon and Dorset, flint from the Cretaceous chalk (obtained often in the form of beach pebbles), and alkali-granite (Cornish stone). Most of the literature of these materials is enumerated and discussed by Williamson (1938). The china and ball clays can be assumed to be dominantly kaolinitic.

The flint is calcined before it is incorporated in the "body", this is primarily to make it more friable and thus easier to grind. The calcination causes the flint to show a decrease in refractive index, as determined by immersion, and also in density. Further, the flint, which exhibits aggregate polarisation initially, may become largely isotropic. The production of cristobalite does not seem to be essential to the initiation of any of these changes. Cristobalite is formed, however, if calcination is adequate, but it shows an anomalously low density. It has been suggested that the fall in the density of flint, which becomes obvious at temperatures as low as 300°C., i.e., prior to the formation of cristobalite, is only apparent. It has been held to depend on the development of minute voids in the structure of the flint which persist even in the cristobalite which is ultimately formed (Simpson and Webb, 1939; Holdridge, Nancarrow and Francis, 1942). The tendency to isotropism could be ascribed to the comminution, during calcination, of the crystalline units in the flint (Williamson, 1941). With this phenomenon may be contrasted the recrystallisation of flint to a quartz-mosaic effected by thermal metamorphism in Antrim (Tilley and Harwood, 1931; Tilley and Alderman, 1934). Harker (1939) cites comparable occurrences in the cherts of various localities. These natural effects may have been determined by the more extended heating, perhaps under "wet" conditions, which the flint or chert received.

Before they are mixed with the clays, the calcined flint and the alkali-granite are ground with water, generally in ball mills. It is possible that the student of natural subaqueous abrasion may find interesting data in the literature of wet grinding of which Creyke and Webb (1941) provide a useful bibliography. These workers have indicated that, in their own experimental grinding of a chalcedonic silica, the rate of comminution increased as the thickness of the hydrated shells around the particles decreased. This thickness is controlled by the nature and the amount of the electrolytes present. Grinding is retarded by the presence of clay. All these observations may have significance in analogous processes of natural abrasion.

The four constituents of the earthenware body are delivered separately to the mixing ark as aqueous suspensions. The suspensions of ball and china clays are discharged into the ark before those of flint and alkali-granite. This is believed to impede preferential settling and consequent segregation. Thus an empirical procedure has been developed which is reminiscent of the experimental results obtained by certain students of the physical chemistry of sedimentation. Boswell (1930) made the significant observation that an apparently even-grained clay may separate into fractions of markedly different grain-size if it is shaken with water and allowed to settle; further, he was able to show in the laboratory that sand or silt could be trapped in a settling, flocculated, clay and size-grading thus prevented. He used colloidal ferric hydroxide to coagulate the clay; Dreveskracht and Thiel (1941) obtained similar results by employing metallic ions as coagulants. The clays used in mixing an earthenware body approximate to the flocculated state. It would thus appear that the potter, by running the clay suspensions into the mixing ark before the addition of the coarser grained flint and alkali-granite, has, unconsciously, forestalled these experimenters in certain of their results and has, moreover, obtained conditions which in Nature would impede size-grading in a sediment.

The aqueous suspension of "body" is now "de-watered" in the filter-press to obtain a plastic mass This process involves the formation of a cake against filtering cloths which retain solids while permitting water to pass. The particles in the cake tend to lie with their major surfaces parallel to the surfaces of the filter-cloth (Williamson, 1941). Identical arrangements of tabular particles can be produced by gravitational settling and can be matched by the disposition of the minerals in some igneous bodies, e.g., the gabbro of the Bushveld Complex (Van den Berg, 1946). In the ceramic filter-press, direct mechanical squeezing of the press-cakes does not occur. Pressure is exerted, however, on the solidifying cakes via the liquid column from the pump; thus pumping at an increased pressure produces a filter-cake with a reduced water-content. This squeezing out of water from a mush resembles the mechanical expulsion of residual melt, entrained by crystals, which has been postulated as a factor in the differentiation of magmas. Williamson (in discussion on Van den Berg, 1946) suggested that the linear parallel structures in the Bushveld gabbro might depend on the movements of entrained liquid consequent upon such expulsion, while the banding in the gabbro might be analogous to the segregation-phenomena that are sometimes exhibited by ceramic filter-press cakes.

The earthenware cakes are now fed to a pugmill in which they are cut by knives, compressed, and forced through a die. The resulting extruded column has structures similar to those found, for example, in the Drachenfels volcanic plug (Williamson, 1941). The particles towards the outside of the column show a planar parallelism that follows the surface. Interiorly, however, there may be structures resembling the flow-lines, produced by tension, which cross the head of a column of magma which was moving forwards before solidification. The analogies between the structures in a column of pugged clay and those in an igneous plug or comparable intrusion could be anticipated from the laboratory studies of Riedel (Williamson, op. cit.), or indeed from the much earlier demonstrations of Daubrée (1879; figs. 141 and 142), in which plastic clay was forced through an orifice. Further experimental work may demonstrate that correspondences exist between the arrangement of joint-systems in igneous rocks and the pattern of drying-cracks that appears in shapes fashioned by the deformation of plastic clay. Certain small fissures in clay-discs have already been shown to be similar in disposition to the fans of tension-cracks which cross the flow-lines in certain igneous intrusions (Williamson 1947). These fissures are not, indeed, drying-cracks but it is to be anticipated that such cracks could develop preferentially at right-angles to directions of tension induced during the fashioning of a ceramic object.

During the biscuit-firing the body attains temperatures of ca. 1150-1200°C. in 60-80 hours; only a few hours are spent at the peak temperature. The effects of firing on the body-constituents may be summarised thus:—

(1) The clay loses chemically-held water; the process becomes marked at ca. 500°C. The solid residue may be regarded as a mixture of finely-divided silica and alumina, although some measure of combination between these oxides has also been considered (cf. Mellor, 1938). As the temperature rises mullite and cristobalite are formed.

(2) The flint changes progressively to cristobalite, although tridymite is the stable phase in the prevailing temperature-range.

(3) The feldspar forms a viscous melt; this melt, during the cooling of the ware, congeals to a glass which serves to bond the new and residual crystalline phases together. Recognisable leucite, produced by the incongruent

melting of potash-feldspar, is unlikely to be encountered because of the brevity of the thermal treatment (cf. Morey and Bowen, 1922).

The mullite and cristobalite in the fired ware are usually insufficiently individualised for discrimination with the polarising microscope; X-ray diffraction patterns and, for cristobalite, thermal expansion data also, may indicate their presence. Quartz, derived largely from the alkali-granite, and flint (cristobalite in part) which is largely isotropic, are very obvious. It is probable that the feldspar-melt tends to dissolve the fine-grained silica made available by the decomposition of the clay, in preference to the coarser quartz or flint (McCaughey, 1937). Bowen (1943) also has indicated a mechanism by which free silica may persist in whiteware to high temperatures.

Mica, from the alkali-granite, may be represented by birefringent pseudomorphs of unknown nature. These, together with the grains of free silica, may, by the arrangement of their major axes, indicate the structural parallelism impressed on the "body" during the plastic deformation that occurred when the articles were shaped; the type of gravitational settling that accompanies the making of ware by casting also produces a specific fabric (Williamson, 1941). The fired ware thus presents a type of "palimpsest" structure.

According to Harker (1939) the presence of solvents is very important for the furtherance of metamorphic changes; these solvents include water. Their actual quantity may be very small. The solvents are believed to induce, in some circumstances, the formation of local glassy patches out of which new minerals crystallise. The solvents, initially held by the glass, are released during the process of crystallisation to perform their work elsewhere. In repetitions of the consecutive processes of glass-formation and crystallisation a small amount of solvent can effect a piecemeal reconstitution of the rock.

In the reconstitution of an earthenware body that occurs during firing the conditions, on the other hand, would appear to be "dry." Further consideration, however, lends less support to this obvious deduction. It is significant that "overfired" bodies "bloat" or develop vesicular structures that can result only from the liberation of gas. According to Austin, Nunes, and Sullivan (1942) such gas cannot be regarded merely as air that was trapped initially in the pores of the unfired body. Bennett, Dawson, Wooliscroft, and Webb (1941) have investigated the "bloating" of fused granite (Cornish stone) and bodies containing it. They direct attention to the observations of themselves and others that the evolution of water from silicate melts can still occur even at high temperatures. It may be noted that Larsen and Switzer (1939) found 4.57% of water in the glass formed by the natural fusion of a granodiorite by an andesite, although they considered that some of this water had been contributed by the andesite magma. Further, buchites, which earthenware so much resembles, may be appreciably hydrous (Thomas, 1922; Harker, 1939). It may be argued that this hydration entirely post-dates the formation of the buchite. It is noteworthy, however, that buchites may be vesicular (Thomas, 1922) and thus resemble a "bloated"ceramic body. The vesicular character indicates the tendency for a volatile phase to disengage itself from the melted rock. Again, the naturally fused granodiorite described by Knopf (1938) is significantly vesicular and can be directly compared with the artificially bloated granite of Webb and his co-workers (op. cit.). It must, of course, be borne in mind that the solution of volatiles in naturally fused rocks and the vesicularity that arises from their release may depend on the application and subsequent relief of pressure. This may well be true of the buchites described by Thomas (op. cit.), which appear to have been carried up from greater depths. In fused ceramic glasses alterations in pressure are much less likely to be involved.

We are thus left with the impression that, during the thermal reconstitution of some ceramic bodies, volatile solvents may be present, but we cannot assess how far, if at all, they assist in this reconstitution.

The "fired" earthenware body shows a curious property; it expands permanently by a process of hydration. This "moisture-expansion" is among the factors that contribute to the formation of minute tension-faults or "crazes" in the glaze. Comparable "moisture-expansion" has been recorded of a volcanic tuff containing glass and quartz-fragments, which is thus not unlike an earthenware body in constitution (Maloy and Lowe, 1945).

The mechanics of moisture-expansion are little understood. Both chemical and physical mechanisms could be postulated. Among the possible physical mechanisms is an alteration in the balance of the surface and elastic forces that give rigidity to a solid; this alteration would follow the imbibition of water (vide H. W. Webb, in discussion on Bangham, 1947B).

MAGNETITE IN FIRED CERAMIC MATERIALS

Thellier (1933) has noticed that fired ceramic articles, which have cooled in a magnetic field, exhibit a weak permanent magnetism. The observation was not novel, for Mellor (1914B) had shown that materials as diverse as bone china and common brick contained crystals of magnetite to which the magnetism could be related. Fe_3O_4 is usually the stable oxide of iron at the peak temperatures of the ovens and kilns of the clay industries. Harker (1939) cites comparable instances of the production of magnetite by thermal metamorphism; among them is the development of this mineral from the Harz hematites in the aureole of the Brocken granite.

CERAMIC GLAZES AND NATURAL GLASSES

The chemical compositions of most ceramic glazes depart widely from those of rocks. Structural analogies may be expected, however, between glazes and volcanic glasses; indeed, bubbles, flow-lines, and spherulitic and other crystal-arrangements resulting from devitrification, are common to both.

The composition and cooling-rate of most glazes are so adjusted that devitrification is prevented and a bright glaze results. In "crystalline" glazes, however, devitrification is deliberately promoted; Norton (1937) has made a careful study of the conditions necessary for the controlled production of these glazes. A well-known type of glaze contains zinc in amounts sufficient to yield megascopic crystals of a compound that Mellor (1914A), on the basis of a quantitative chemical analysis, has identified as willemite. Norton (op. cit.) appears to be unsatisfied by this identification. The compound has the interesting property of taking up cobalt from the surrounding glaze. The crystals then become blue and the glaze in this vicinity is correspondingly paler (Mellor, 1937, provided a good colour photograph). Singer (1946) suggested that the similarity in the ionic radii of Co·· and Zn·· was the basis of this phenomenon; further, he stated that if cobalt and manganese co-exist in a glaze, the zinc silicate crystals abstract the cobalt preferentially. Thus the cobalt colours the crystals and the manganese the groundmass of the glaze. The impoverishment of a glaze in cobalt in the vicinity of zinc-containing crystals appears to be a specific example of the development of the "crystallisation haloes" that are not infrequent in partially crystallised glasses. Thus such haloes are reported and figured around hercynite in the glassy products of the reaction between the contents of a glass-furnace and its refractory lining (Seal, 1948) and around magnetite of similar occurrence (G. V. Wilson, 1918). Comparable examples of the "clearing" of glass around crystals contained therein are to be found in the pitchstones of Arran (Tyrrell, 1928) and in the glass of a naturally fused granodiorite described by Larsen and Switzer (1939).

THE PETROGRAPHY OF SOME TYPICAL REFRACTORIES

Firebricks

The commonest type of firebrick is made from a clay that contains free quartz. A well-burnt brick is likely to show mullite and cristobalite, derived from the clay-minerals, with some residual quartz. The crystalline phases are cemented by a glassy bond. More intensive thermal treatment of the brick causes enhanced solution of the free silica in the fused glass (e.g., see Jay, 1939).

Silica Bricks

A typical silica brick contains, when still unburnt, quartzite fragments in a matrix of slaked lime and quartz-flour. If the temperature and duration of burning are adequate, the quartzite fragments change wholly, or more usually partly, to cristobalite. This change begins at the exteriors of the fragments and along the intergranular or other cracks that penetrate them. The matrix of the fragments is, on the other hand, a meshwork of tridymite units in a lime-rich glass that may contain some crystals of a pyroxene, such as wollastonite. Quenching experiments by Scott (1925-26; 1930) suggest that this tridymite meshwork is in existence at high temperatures and does not make its first appearance only during the actual cooling of the brick. The lime-rich matrix thus serves as a mineralising agent. Tridymite may be found also as fringes to the cristobalite-quartz assemblage which represents the original quartzite fragments (Thomas, Hallimond, and Radley, 1920; Rigby, White, Booth and Green, 1946).

It has long been realised (e.g., by Thomas, 1922) that silica bricks and siliceous xenoliths resemble each other in both mineralogical constitution and structure. Thus the thermal metamorphism of a feldspathic sandstone may yield a rock containing quartz fragments in a glassy matrix from which tridymite has crystallised (e.g., see Harker, 1939, fig. 21A). However, cristobalite is not reported in association with the quartz fragments. The tridymite is commonly replaced by pseudomorphous quartz; in silica bricks, on the contrary, the tridymite is preserved. The rapid cooling of silica bricks after burning may be responsible for this difference. Further, we do not know if the tridymite would change to quartz when bricks were stored for periods of time geological in duration.

Lacroix (1946) also has described siliceous xenoliths and compared their structures with those of silica bricks. However, he made the fantastic assumption that the appearance of glass around the quartz of the xenoliths indicated that temperatures of at least 1750°C. had been attained. It is clear from the investigations of Holmes (1936) and of Holmes and Harwood (1936) that such glass cannot be assumed to be fused silica; indeed, some glass described by these authors had the composition of potash-rich acid obsidian. Lacroix offers no chemical data in support of his thesis; the fact that the glass he described contained crystals of pyroxene suffices to indicate that it is unlikely to be merely fused silica.

Magnesite Bricks

Although these bricks are named after the raw material from which they are manufactured, they really contain periclase derived from it by calcination. This periclase may be accompanied by minor amounts of other phases of which the nature is determined largely by the types and amounts of impurities in the initial magnesite (cf. Rigby, Richardson and Ball, 1947).

In one manufacturing process, periclase, produced from magnesite by calcination, is allowed to undergo partial rehydration; brucite is formed (Chesters and Weyl, 1932-33). Thus the synthesis of brucite follows the mechanism that has been postulated for the formation of the mineral in thermally metamorphosed magnesian limestones. This classical explanation of the formation of brucite in such limestone has, however, been rejected by Keith (1946) for certain specific occurrences in Ontario; he prefers an hypothesis of the direct replacement of dolomite through the agency of magnesia-bearing gases or solutions. Tilley (1947) has, however, recently mentioned an occurrence where the classical theory can be invoked.

Calcined Dolomite

It has been shown experimentally that the initial change during the calcination of dolomite involves the formation of periclase and calcium carbonate (Bradley and Jay, 1946); with this result may be compared the observed occurrence of periclase and calcite in thermally altered dolomites or dolomitic limestones. A difficulty encountered in the use of dolomite refractories that contain some silica is the tendency to "dusting" which depends on the change from beta to gamma $2CaO.SiO_2$. The same phenomenon is seen in contact metamorphic assemblages containing $2CaO.SiO_2$; it may be obvious in the field (Tilley, 1947) or become apparent when material is ground during the preparation of thin sections (Tilley, 1929).

Chromite Bricks

These are made by "burning" chromite with its associated gangue minerals such as serpentine. The matrix of the brick develops from the gangue and additions are sometimes made to increase the refractoriness of this matrix. The writer is unaware of any record of the natural thermal metamorphism of a chromite deposit that would permit a comparison to be drawn. The study of chromite by the ceramist has, however, yielded results that may interest the mineralogist. This study has involved the investigation of natural chromite, the synthesis of a number of spinels, and the investigation of the effects of heating and of various atmospheres upon natural and artificial spinels (see especially Green et alia, 1946).

Chrome-Magnesite Bricks

The prominent constituents of the burnt bricks are periclase, forsterite, and chromite.

CARBON DEPOSITS IN FIREBRICKS: THEIR ANALOGY WITH NATURAL GRAPHITE

Fragments of iron-rich minerals may exist in fireclay and, in bricks made from it, become loci of carbon deposition. This deposition occurs if the bricks are used in an atmosphere, for example that of the blast furnace, containing gaseous compounds of carbon. The loci are associated with grains of magnetite or other ferruginous material which promote the formation of graphite from carbon monoxide or gaseous hydrocarbons. The brick is actually disintegrated by the formation of graphite in its pores.

The occurrence of this type of reaction should be of interest to those who have suggested that volatile compounds of carbon are involved in the formation of deposits of natural graphite. Davidson (1942-44), for instance, could accept carbon disulphide or oxysulphide, or carbon monoxide or dioxide, as the source of the graphite in certain crystalline limestones of the Outer Hebrides. Here dedolomitisation could have made carbon dioxide available. Similarly Wadia (1943) suggested that carbon monoxide or dioxide were involved in the formation of the graphite deposits of Ceylon. Dissolution of limestone in charnockite magma could provide the carbon compounds.

THE ATTACK OF MOLTEN SILICATES ON REFRACTORIES

Most refractories are not homogeneous; they consist of grains of high melting or softening point with, in the intergranular spaces, more fusible material which, in some instances, is the glass that developed when the refractory article was burnt during manufacture. Attack by a silicate melt is thus likely to be preferential. Pores are invaded and glass or other readily fusible constituents assimilated by the melt; particles of high melting point or the refractory reaction-products of these, tend to pass into the aggressive liquid as slowly dissolving xenoliths or xenocrysts. Thus Partridge and Seal (1944) discussed the implications of the well-known observation that molten glass enters glass furnace tank blocks via the intergranular glassy layers and releases "stones" which float away in the fused batch. These "stones" or xenoliths may consist of grog (pre-calcined clay), sillimanite, mullite, etc., or their reaction-products such as corundum. Comparable phenomena are demonstrable at igneous contacts; Williamson (1935), for instance, described the attack of a granodiorite magma on a gneiss which had been thermally altered to alternating quartz-feldspar and aluminous hornfels layers. The less refractory quartz-feldspar layers were penetrated in preference to those of hornfels.

The attack of fused silicates on refractories is likely to be enhanced if compounds capable of providing low-melting eutectics can be formed with some of the material attacked, although the rate of reaction and the "wettability" of the refractory have also to be considered (Norton, 1942, pp. 494-521). Thus the volume of the attacking melt is augmented and the phases most resistant to solution are launched therein. Brammall and Harwood (1932) deduced just such a mechanism for the interaction of the Dartmoor granite and the shales of its aureole. Silica and alkalies from the magma promoted the formation of quartz and feldspar in the shale; the selective fusion of these minerals allowed andalusite and other refractory compounds that had developed in the shale to pass into the melt as solid phases. In this instance only some constituents of the magma entered the shale. Similarly certain constituents of an invasive silicate melt may effect preferential entry into a refractory with which the melt is in contact; Seal (1948) cites an example where Na_2O and B_2O_3 from molten glass behaved in this way.

The attack of silicate-melts on refractories or of igneous rocks on their confining walls, if it follows the schemes just outlined, bears a resemblance to the marine erosion of a series of alternating hard and soft sediments. However, the attack may achieve more complicated results by the production of selvedges, skarn-like deposits, and zonary structures. Thus a selvedge accompanies the Dartmoor granite contact just discussed. Artificial selvedges and cognate structures are described later.

Preston and Turnbull (1941) have discussed the mechanism of a peculiar form of attack by molten glass on refractories known as "upward drilling." This attack is characterised by the appearance of pits on the downwardly facing surface of the material affected. These pits mark the sites of bubbles in the glass and the enhanced corrosion of which they are witnesses depended on periodic changes in the surface energy of the melt bounding the bubbles. Comparable natural phenomena do not appear to have been recognised, but Preston and Turnbull have suggested that geologists should consider if volcanic pipes could originate by "upward drilling."

CONTACT AND RELATED PHENOMENA IN GLASS FURNACES

The Classical Investigation by G. V. Wilson

G. V. Wilson (1918) made a notable contribution to the study of those phenomena of interest to the geologist that may be encountered on glass works. He had, for instance, access to a burst tank furnace from which the molten glass had poured into the arched space beneath. Here the glass, by entering into various relationships with the bricks and mortar, had simulated the structures of an igneous complex. These structures varied from injection breccia, through simple and composite veins, to stringers containing both glassy and crystalline material where the glass was disposed either marginally or centrally. Various artificial minerals were also to be found; of these perhaps the most unusual were euhedral quartz and a species identified as biotite. The premature chilling of the glass had left for study xenoliths of limestone that, in circumstances more propitious to the glass-maker but less fortunate for the mineralogist, would have dissolved completely in the fused batch. These xenoliths had paragenetic mineral-associations resembling those of the chalk-dolerite contacts of Scawt Hill, Co. Antrim, Ireland. Associated with the xenoliths were augite, wollastonite, melilite, and the compound $3CaO.2SiO_2$. The mineral form of the latter has recently been identified by Tilley (1942) at Scawt Hill and named "rankinite." The same authority had already shown that $2CaO.SiO_2$ occurred naturally in the same locality. This mineral he christened "larnite" (Tilley, 1929). He has recently demonstrated, with Vincent, that, in addition to β-$2CaO.SiO_2$ or larnite, a rarer mineral, which is α'-$2CaO.SiO_2$, is to be found; for this the name "bredigite" was proposed (Tilley and Vincent, 1948). Further, Tilley (1942) has identified larnite in Wilson's artificial material. In both the natural and artificial assemblages rankinite separates wollastonite from larnite thus—

Locality	*Source of Lime*	*Artificial Compound or Natural Mineral*			*Source of Silica*
Glass Furnace	Limestone Xenolith	$2CaO\ SiO_2$	$3CaO.2SiO_2$	$CaO.SiO_2$	Molten glass
Scawt Hill	Chalk	Larnite	Rankinite	Wollastonite	Flint nodule

It is interesting to note the similarity of this scheme to that used by Jander and Hoffman (1934) to illustrate the results of their studies of the reaction between CaO and SiO_2 in the solid state at temperatures from 1000°C. to 1200°C. Their schematic figure No. 8(3) shows the same calcium silicates as are indicated in the above scheme in precisely the same order between solid lime on the one side and solid silica on the other. They believe that CaO diffuses more rapidly than SiO_2 and determines the arrangement of the calcium silicates between the two solid reactants. Bowen (1940) has suggested the phases that may be expected to come into existence in a siliceous limestone as temperature rises; these phases have a definite and progressive sequence. In this sequence wollastonite appears earlier than larnite, and Bowen suggested that rankinite, the supposed absence of which at Scawt Hill he comments upon, has a minimum temperature of existence above the maximum temperature prevalent during thermal metamorphism. The arrangement of the calcium silicates at Scawt Hill and in Wilson's material, together with the experimental studies of Jander and Hoffmann (op. cit.), indicate that diffusion rather than thermal stability may determine the nature of the calcium silicate produced. Indeed, Jander and Hoffmann state that $2CaO.SiO_2$ is in general the primary product of the lime-silica solid reaction. No doubt, however, there may be time-temperature conditions where, as assumed by Bowen, thermal stability is of major importance in deciding which calcium silicate shall appear, or more probably, survive.

The Formation of Alkali Minerals

The presence of abundant alkalis in many glass batches allows the synthesis of alkali-minerals. Thus Lambeth (1946) has described the formation of nepheline and aegirine in the fireclay recuperator tubes of a glass furnace. The minerals originated from soda out of the glass batch and Fe, Al, and Si contributed as fly ash by the gasified coal in the producer. The fireclay refractories were therefore not involved. McCaughey (1937), however, has illustrated a reaction, in a similar environment, in which highly aluminous checker bricks have, with sodium sulphide, yielded lapis lazuli; the sodium sulphide is made available by the reduction of sodium sulphate in the glass batch.

Most glass tank furnace blocks contain appreciable mullite; during the use of such blocks the mullite may become involved in reactions that produce corundum and alkali-minerals. For example, Insley (1945) has mentioned that, above the glass-line, alkali-vapours and dust may convert the blocks to a meshwork of corundum crystals, from the interstices of which a molten alkali-containing silicate may drain. Seal (1948) has commented on the fact that the presence of foreign materials allows the meritectic reaction that produces corundum from mullite to occur at lower temperatures. Corundum so derived may enter the molten glass to give "stones," while the invasive foreign materials unite with the residue of the mullite molecule to provide a liquid which drains from the tank block and is capable of crystallising as oligoclase, labradorite or, more rarely, nepheline (Insley, 1945). The course of the reaction below the glass-line, i.e., where molten glass itself is available, is in general similar. Insley notes an increase in the amount of glass in the refractory, the development of corundum, and the formation of a melt which, in appropriate circumstances, may crystallise as nepheline and carnegieite. Even a refractory containing appreciable free quartz can yield corundum in this way. The chemical mechanism producing corundum is reminiscent of the desilication-process which Du Toit believed to be responsible for certain natural occurrences of this mineral (vide Du Toit in discussion on Brandt, 1946). Insley, however, directs attention to analogies between the artificial mineral-assemblages containing corundum and certain natural assemblages that differ somewhat from the desilicated pegmatites studied by Du Toit. These natural assemblages are the associations of corundum with syenite or nepheline syenite at contacts with alumina-containing country-rocks. Such contacts can be likened to those between molten glass and aluminous refractories. Because tank-block compositions containing appreciable free quartz are still able to yield crystallised alumina, Insley surmises that the country-rock associated with a syenitic intrusion need not be abnormally aluminous in order to produce corundum.

THE SYNTHESIS OF ALKALI-MINERALS IN THE BLAST FURNACE

Alkali-vapours from the burden are known to cause the formation of alkali-minerals in the firebrick of the blast-furnace. Rigby and Richardson (1947) have shown that kalsilite and leucite are so produced, apparently by a type of artificial pneumatolysis or metasomatism. Such occurrences recall the demonstration by Holmes (1945) of the leucitization of certain rocks by potash-rich lavas; the rocks so affected were granite and, to quote a rock nearer in chemical composition to the firebrick of blast furnace linings, phyllite.

SKARN FORMATION AND COGNATE PHENOMENA

The contact between igneous and country rock is sometimes accompanied by a zone in which one or more mineral species are concentrated. These species may or may not occur also in the invaded country rock. Thus Williamson (1935; fig. 8) depicts zones of hornblende or pyroxene at the contacts between igneous veins and hornblendic or pyroxenic hornsfels respectively. Nockolds (1934) instances remarkably similar zones and applies to them the term "skarns." The examples of skarns just cited are very small-scale phenomena; the choice of examples has been made to facilitate comparison with ceramic occurrences of not very dissimilar magnitude. Thus a type of skarn appears at the interface between a glaze and a pottery body if the glost temperature has been sufficiently high. Mellor (1907) long ago provided a photomicrograph of such an occurrence in a crucible of Berlin porcelain; acicular crystals project into the glaze from the surface of the body. Similar occurrences are common and, where such fringes of acicular crystals exist, the surface of the associated body has usually been markedly corroded by the molten glaze. The crystals are commonly identified as mullite. This compound, which may exist in the body itself, may be considered to have grown under the mineralizing influence of the glaze. The solution of body-constituents in the glaze, followed by the crystallisation of these materials as the glaze cooled, also can be regarded as having increased, or even initiated, the growth of the mullite fringe. Some skarns at natural contacts could owe their formation to the regrowth of minerals pre-existent in the country-rock; other skarns contain mineral-species not to be found therein. Artificial skarns show the same phenomena. While some consist of mullite, which is seen also in the body, others contain compounds which the body does not. Such compounds were noticed, for instance, at the glaze-body interface, by Parmelee and Buckles (1942); anorthite was among them. Indeed, anorthite and other plagioclase feldspars are common reaction-products also at contacts between glasses or slags and refractory materials. G. V. Wilson (1918) described oligoclase which appeared at the junction between glass and a brick. Towards the brick the oligoclase was rich in "sillimanite"; this observation was held to indicate that the brick had been supplying alumina for the synthesis of the feldspar. The feldspathic reaction-rims surrounding some of the xenoliths in the tholeiites of Mull show a similar concentration of "sillimanite" towards the buchites which they border (Thomas, 1922). Phemister (1938-42) noticed a bytownite skarn between a slag and a firebrick. Hugill and Green (1939) provide remarkable serial photomicrographs in which the genesis of anorthite-skarn, by reaction between blast furnace slag and firebrick, can be traced.

Some type of diffusion must be involved in the production of these feldspathic artificial skarns. Biotitic selvedges, which are among their natural analogues, have been ascribed by Reynolds (1947) to the operation of diffusion. Solid diffusion, however, is specifically implied; this type of diffusion cannot be the main mechanism involved in the formation of artificial skarns, for here molten material is available. The problem of solid diffusion in ceramic materials receives further attention later.

ZONARY STRUCTURES IN SLAGGED REFRACTORIES

Special interest is attached to these structures because, if it is permissible to borrow the terms from petrology, they provide evidence of "fronts" and "culminations" as defined by Reynolds (1946).

The classical example of a zonary structure appears in used silica bricks from the roof of the basic open hearth steel furnace. Over 20 years ago Scott (1925-26) gave a careful description of the phenomenon. The considerable literature that has now accumulated was recently critically appraised by Dodd and Green (1939). The invasive ferruginous material from the furnace constitutes an iron "front"; as this front is established the lime-silica matrix of the brick retreats and thus gives rise to a lime "front" a few inches behind the hot face of the brick. The existence of the lime "front" is witnessed, in the cold brick, by the presence of a zone rich in yellowish glass. In this zone lime, and also alumina, achieve a "culmination." Scott (op. cit.), Greig (1927), and later Dodd and Green (op. cit.) have indicated that the location of this culmination is determined by a correlation between the fusion temperatures of the relevant mixtures and the temperature gradient in the brick. The effect of the temperature gradient on the disposition of the zones, relative to the hot face, is readily demonstrated from the data that Harvey (1935) has presented, partly in the form of colour photographs; the disposition of the zones is seen to have been altered by backing the brick with heat-insulating material and thus modifying the temperature-distribution within it.

The invasion of chrome-magnesite bricks by ferruginous material produces a zonary structure which is well described by Hugill and Green (1938, 1939); this structure appears to arise by a mechanism somewhat different from that which operates in silica bricks. The hot face of the attacked brick develops a blue-black, loosely granular, and apparently unfused layer, but behind this is a brown layer showing signs of fusion. It is generally agreed that the incoming iron causes a swelling of the chromite grains and thus their magnesian matrix is compelled to migrate away from the hot face. The solidification of this migratory material produces the brown layer showing signs of fusion; in this layer there is a "culmination" of magnesium. Zerfoss and Davis (1946) have suggested a mechanism by which the migration of magnesium could be assisted; forsterite, which forms the matrix of the chromite at the hot end of the brick, is assumed to take up iron and thus to become more fusible. The iron-enriched olivine travels towards the interior of the brick where iron is abstracted from it by periclase and chromite; its solidification as a more magnesian, i.e., more refractory, olivine is thus assisted.

Zerfoss and Davis (op. cit.) consider that the swelling of the chromite-grains, by virtue of which the matrix of the brick is constrained to move, depends on the deposition of magnetite on chromite-nuclei; indeed, the grains of chromite become demonstrably euhedral as their re-growth proceeds (Hensler and Zerfoss, 1947). Zerfoss and Davis (op. cit.) admit, however, that actual solid solutions between chromite and magnetite may form. This admission accords with the change in lattice spacing, noted by Lynam, Chesters, Howie and Jay (1942), which occurs when chromite absorbs magnetite. The work of Lovell, Rigby and Green (1946) has, however, indicated the complexities of the problem of the "swelling" of chromite-grains; they suggest that "bursting expansion" occurs if an actual structural re-arrangement of the spinel lattice accompanies solid solution.

Adams (1930) has described certain artificially produced zonary structures; his description has since been quoted on several occasions in petrological literature. An edge of a dolomitic magnesite brick, containing magnetite, was in contact, at temperatures up to ca. 1430°C., with the face of a low lime magnesite brick. Although the latter brick retained its shape, it showed, in the vicinity of the line of contact, an enrichment in iron and magnesia; the iron was certainly invasive. The incoming material caused a lime-alumina-silica front to develop ahead of it, reminiscent of that found in silica bricks from the roof of the basic steel furnace. Thus lime, for instance, rises to a "culmination" and then falls sharply. Read (1934) remarked on the similarity between the phenomenon described by Adams and certain zonal associations of minerals that appeared around isolated fragments of ultra-basic rock when the country rock was permeated by acid materials. The oxides in these zones showed typical "culminations." Phillips and Hess (1936) described comparable zonal associations of minerals and, independently, postulated processes of migration to account for them which were similar to those which Read assumed to have occurred; however, they dissented from Read's opinion that the various zones had been formed contemporaneously.

THE PROBLEM OF DIFFUSION AND REACTION IN THE SOLID STATE

The hypothesis that diffusion and reaction in the solid state, without the intervention of a liquid or gaseous phase, are involved in petrogenesis has lately been resurrected from obscurity. Only a few years ago it was regarded with open disfavour. Thus Harker (1939) stressed the importance of solvents in promoting reaction and repeated the dogmatic statement, already made in the first edition of his book (1932) that "no sensible reaction can in general be verified at the contact of two crystalline bodies." This may well be true of those instances where, as he points out, a thin section of a metamorphic rock may show several bands of contrasting chemical composition in a single field of the microscope. The statement, however, receives no support from the experimental results of Hedvall and many other investigators (Hedvall, 1938A, 1938B, 1948). Nevertheless, it can be granted that reactions occurring by virtue of solid diffusion may be accelerated by the presence of water-vapour, e.g., the reaction between calcium oxide and silica studied by Jander and Hoffmann (1934) behaves in this way. Read (1933), writing at about the same time as Harker, criticised Stillwell for laying too much stress on solid diffusion as a factor in the genesis of metamorphic rocks; he believed that many petrologists would join him in this criticism. In 1939 he still regarded solvents, including water, as necessary to the changes accompanying regional metamorphism (Read, 1939). Some recent publications incline to a different view. Thus Holser (1947) clearly recognised that diffusion of ions, in the absence of a fluid, may be of great importance in metasomatic processes, while Bugge (1946) laid great stress upon solid diffusion as a promoter of metamorphic changes, particularly at depths where intergranular penetration by fluids may be expected to be difficult. Holmes and Reynolds (1947), discussing metasomatic metamorphism in Donegal, cited evidence which suggests that migration of ions along crystal boundaries and through crystal lattices, rather than the intervention of liquid and gaseous phases, have produced the phenomena which they describe. Barth (1947) is prepared even to accept the formation of some rocks by petroblastesis, i.e., by the direct crystallisation of diffusible ions; if the ions do not crystallise per se, but react with a pre-existing rock, the process is still classed by Barth as one of metasomation (cf. Holmes and Reynolds, op. cit.). Backlund (1946) points out that the somewhat elusive emanations that are thought by certain petrologists to be potent factors in petrogenesis should be regarded as migrating ions rather than as liquid or gaseous media for ionic transport. However, the two conceptions of emanations are not necessarily sharply distinguishable, for the ions at crystal interfaces are in an exceptional condition and capable of travelling virtually as a fluid film.

It is thus worthwhile to review ceramic phenomena to see if they provide evidence of solid diffusion and reaction which, by virtue of analogies with inferred petrogenetic mechanisms, may be of interest to the geologist. Reynolds (1942), for instance, directed attention to the zoned magnesite brick described by Adams (1930) and discussed earlier in the present study. Reynolds stated—"if it be objected that migration of elements or oxides is an unknown physical process, reference need only be made to the solid diffusion in bricks described by Adams (1930)." Backlund (1946) appeared to interpret the results of Adams in a similar way. Actually there is no evidence that solid diffusion was the principal mechanism involved in the production of the phenomena noted by Adams, although such diffusion may have operated within individual crystals; the lack of microscopical data impedes interpretation. It is probable that an intergranular liquid, determined by the presence of components that could yield low-melting eutectics, was the principal cause of the zonary arrangement. This possibility is explicitly recognised by Adams (op. cit.) and is rendered very plausible by the existence of a region of actual vitrification in the zoned brick near the line of contact with its neighbour. Greig (1927) and Dodd and Green (1939) postulate that a migrating liquid is responsible for the comparable zonary structures in silica bricks. These authors indicate that the concentration of compounds in specific zones follows from the fusion points in the relevant systems and the temperature gradient through the brick (cf. Scott, 1925-26). The fact that a zoned brick, such as the example described by Adams, retains sufficient cohesion at high temperatures to keep its shape, is no argument against the existence of fluid constituents therein. For instance, ordinary white earthenware contains considerable glass, which represents a silicate-melt that once existed in the heated articles, yet it does not become distorted if the "firing" operation proceeds normally.

It has been mentioned already that Read (1934) draws comparisons between certain zoned structures produced by metamorphic differentiation and the phenomena discussed by Adams; he joins Phillips and Hess (1936) in believing that these structures were produced under "wet" conditions. Reynolds (op. cit.) cites the same phenomena to illustrate solid diffusion, i.e., diffusion under "dry" conditions.

Thus the zoned ceramic structures discussed above would appear to depend for their formation more on the existence of an intergranular or pore "magma" than on solid diffusion. However, some ceramic phenomena permit the operation of solid diffusion to be inferred with reasonable certainty. This is true, for instance, of the interaction between aggressive ferruginous material and the chromite of chrome-magnesite bricks (vide supra). The descriptions and photomicrographs of Hugill and Green (1938; 1939) show that the grains of chromite are not merely enlarged by the deposition of magnetite upon them; the original colour of the grains is progressively obliterated by the extension of an invasive darker periphery. Finally even the centres of the grains are affected. Hedvall's account of the diffusion of iron into quartz illustrates a comparable occurrence and one which is on the same small scale (Hedvall, 1938A, 1938B).

The zonary structures found in certain ceramic objects may require, for their formation, migrations a few inches in extent, but even these migrations are not governed by the process of solid diffusion; where the existence of solid diffusion can be substantiated it has a scope which does not go beyond that which is involved, for example, in the exsolutional phenomena shown by feldspars. Some hypothetical petrogenetic mechanisms may require migrating ions to travel for miles by a process of solid diffusion. Geological time is, relatively, enormously long; thus, once we have experimental evidence of solid diffusion, we may argue from the minute to the grandiose, as Backlund (1946) suggests that we perforce must do. However, if we cite artificially induced phenomena to support an argument by analogy, we must first be sure that the analogy is sound.

SUMMARY

Recrystallised "monomineralic" artefacts show analogies to thermally produced quartzites and marbles. Fired complex ceramic bodies resemble buchites; both contain glass and new or residual crystalline phases. Mullite may be one of the new phases. Natural and artificial occurrences of mullite are discussed in the light of recent experimental work. While the alumina-silica compound found in ceramic articles may be referred to mullite with some assurance, there is less certainty as to how much natural "sillimanite" is really mullite. The conversion of kyanite, andalusite, and sillimanite into mullite is important in the manufacture of alumina-silica refractories; the ease of conversion of these minerals, determined experimentally, accords with their behaviour during contact metamorphism.

Both thermally metamorphosed rocks and fired ceramic articles may show a progression towards high temperature equilibria which are not ultimately attained; the phase assemblage that persists after cooling is metastable but indicates how far this progression has gone. It is not certain that the mineralogical changes that accompany the firing of ceramics have always occurred under strictly "dry" conditions. The high temperature "bloating" of some ceramic bodies is significant and recalls the vesicularity of buchites.

The mineralogy of typical refractories is described. Silica bricks resemble siliceous xenoliths. Experimental studies of basic refractories have shown that the formation of periclase from magnesite and, in association with calcite, from dolomite, and the subsequent change of periclase to brucite by hydration, occur in a manner comparable with that suggested by the evidence from igneous contacts.

There are analogies between the action of glasses and slags on refractories and that of igneous intrusions on country rocks. In both instances the melt may be augmented by low-melting constituents, while the more refractory compounds, that are residual or produced by reaction, are launched into the melt as xenoliths.

Reactions between glass and refractories may lead to the synthesis of alkali-minerals. Desilication may produce corundum. Blast-furnaces provide a second locus for factitious alkali-minerals.

The formation of skarns and zonary structures at the contacts between silicate-melts and ceramics is described. Fronts and culminations exist, but it is not certain how far they are comparable in origin with their natural analogues. Migration of a melt rather than solid diffusion appears to be involved in their production. When evidence of solid diffusion is available it indicates a very small-scale effect. Because of the importance of the time factor such evidence cannot lead to outright denial of more formidable effects under natural conditions. A warning is given, however, against the misinterpretation of artificial phenomena in the endeavour to elucidate natural occurrences.

The formation of graphite in firebricks from gaseous compounds of carbon may have significance for theories of the origin of natural graphite-deposits.

Phenomena occurring during the wet-mixing of earthenware are of interest to the sedimentary petrologist.

The manipulation of plastic ceramic clays or bodies produces petrofabric arrangements that can be matched in rocks. Structures found in the unfired body may persist as palimpsests in the fired ware.

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REVIEWS

Fear, War and the Atom Bomb; by P. M. S. Blackett. Pp. vii, 244. New York, 1949 (Whittlesey House, McGraw-Hill Book Co.). Two schools of thought may be discerned among physicists who have written in a popular or political vein about the atomic bomb. The first may be called the idealist school. Its numerous partisans were united by the conviction that the atomic event was so enormous, so unique and lethal in its social significance, as to invite an ignoration of all traditional political concerns and to demand the immediate institution of a world government to prevent war. Their efforts have come to naught for two reasons: one was the dissenting chorus of American voices, mostly from workers in the social sciences, which asserted that a neglect of history and of political institutions is unwarrantable and bound to be disastrous; the other was Russia.

The second school, adequately labelled realistic, though it was perhaps in sympathy with the aims of the idealists, tried to make the best of the actual situation and endeavored to integrate the problem of the atomic bomb with the other political problems of our time. It tended to minimize the importance of the bomb as a weapon and contemplated unflinchingly the possibility of an atomic war Blackett is a leader of the realistic school and, furthermore, an apologist for Russia.

Commanding an unusual store of pertinent information, he analyses the facts surrounding the use of the bombs in Japan, the causes for the decision to use them, and the subsequent effects on international policy. He believes that diplomatic, not military expediency dictated the large-scale killing of civilians, implying that, in the minds of our leaders, the Japanese surrender to the Americans had to be secured before the Russian invasion developed too successfully. Because of the uniqueness of these motives, because of the improbability of their recurrence in another war, Blackett seems to feel that a wholesale use of atomic bombs on civilians in the future is not as likely as is often feared.

Another argument prominent throughout the book bears upon the Russian refusal to adopt the American (Baruch) plan for the control of atomic weapons. The key to an understanding of this attitude, Blackett thinks, is in the need Russia has for atomic energy in peaceful pursuits. Her energy output is low and for that reason she has a low standard of living. America, with her high rate of production, can easily afford to propose measures that will impede the development of atomic resources on a world-wide scale, which according to the author the Baruch plan will do. Russia stands to gain far more from non-military applications of recent discoveries than the U. S., and more from non-military uses than from military ones.

Only a very small part of the book is devoted to the author's own proposals, which are interesting but none too effective; for when the reader has reached the last section, where advice is given, his worry about the soundness of the author's daring arguments has become inveterate and his responsiveness is low.

The idealists have said: let us disregard history; let us build anew. The early realists have surveyed the historical facts in their entirety and have based predictions upon them, often oblivious to the limited predictive qualities of present historical analyses. Blackett analyses with care, though with some obvious bias, the *scientific* facts of the situation and assumes that they provide a basis for historical prediction.

Somehow, I find Blackett's errors less interesting and less commendable than the errors of the idealists. HENRY MARGENAU

Quantum Mechanics; by L. I. SCHIFF. Pp. xii, 404. New York 1949 (McGraw-Hill Book Company, $5.50). A textbook suitable for a one-year graduate course in quantum mechanics. The treatment is modern, succinct; the topics are on the whole well chosen and reflect present-day interests in theoretical physics. As compared with other books in this field, its coverage is very complete and reasonably elegant. It is unique in its inclusion of an introduction to quantum electrodynamics, but it achieves this virtue by an omission of other important topics, such as quantum chemistry and the quantum theory of metals. HENRY MARGENAU

Principles of Petroleum Geology; by CECIL G. LALICKER. Pp. xii, 377; 67 tables, 157 figs., 8 pls. New York, 1949 (Appleton-Century-Crofts, Inc., $5.00). This book is apparently designed as a text for a beginning course in petroleum geology. The early chapters sketch a brief history of the subject and give an extended discussion of the physical and chemical properties of petroleum Launching into geological matters the author presents an array of opinions regarding the origin of petroleum and its migration and accumulation. Reservoir rocks and oil structures are discussed, and numerous examples of oil fields, selected by mode of origin, are described. The current geophysical and other discovery methods are presented and explained. The closing chapters give a summary of geological considerations in recovery methods and the ways in which oil and gas properties are evaluated. No formal bibliography is included in the book, but frequent references are made to the literature by means of standard footnotes.

One might question whether petroleum geology has any established principles, Mr. Lalicker's review of its literature leaves one in doubt on this point. In the reviewer's opinion, the book would be strengthened by a more rigorous treatment of the physics and chemistry in the discussion of the properties of petroleum. The voluminous stratigraphic tables of oil fields would be rendered more valuable by indicating the producing horizons in some way on the charts themselves. Similarly, the many structure contour maps, which are adequately located by township or county, would be more useful if small state index maps were added. Notably absent is any mention of the recovery possibilities from oil shale, such as the Green River formation.

The book presents much valuable material, but its value would be enhanced greatly by revision with particular attention directed to the organization.

JOHN E. SANDERS

PUBLICATIONS RECENTLY RECEIVED

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GRANITE AND METASOMATISM

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ABSTRACT All geologists interested in the genetic problem of granite have long agreed that some granites crystallized from their own melts (magmas), while other rocks of closely similar composition and texture were formed by molecular replacement (metasomatism) of older, non-granitic rocks by absorption of foreign material According to the cumulative opinion of most experienced geologists during the past hundred years, the ratio of total volume of "magmatic granite" to total volume of "metasomatic granite" in the earth's crust is large. Within the last twenty-five years a smaller but increasing number of workers, here called for convenience "metasomatists," have come to prefer a reversed ratio. The following paper is intended to show that the second group has not thought "to scale," have neglected many vital facts bearing on the problem, and have in no wise weakened the arguments of the "magmatists," which have led toward what appears to be a sound theory of the evolution of the earth's crust.

INTRODUCTION

POSED more than a century and a half ago, the granite problem is one of peculiar dignity and meaning. If it were solved, the origin of basalt, peridotite, anorthosite, gneiss, and other principal stuffs of the earth's crust would be more intelligently discussed, and theories relating to continental genesis, geophysics, ore genesis, geodesy, physiography, oceanography, and paleontology would be better controlled. If it were solved, man would know with more assurance why there is dry land, a perch for air-breathers like himself; how he can thrive in a metal-hungry age; and how to account for the sedimentary rocks which bear the medals of creation and declare so much of the history of our planet. But there are two general and portentous difficulties in the way when the attempt is made to round out a good theory of granite itself.

First, the solution cannot be complete until the cosmogonists declare the mode of birth, the initial temperature, the initial composition including gas-content, and the youthful organization of the nearest of the heavenly bodies; here geologists lack some of the guiding data they need.

The second lion in the way toward the desired goal is an ineluctable necessity. No solution of the granite problem is possible until there be assembled and digested the relevant facts recorded by polyglot field-workers in seven continents and the wide oceans. But even this world-wide, encyclopedic product of field work with hobnails, hammer, compass, and topographic map is manifestly insufficient. Other data, also widely scattered in the great libraries, must come: (1) from properly designed experiments on melting, molten, and freezing silicates and mixtures of silicates, even if only in "little crucibles"; (2) from analogies secured by experiments with materials other than silicates; (3) from handbooks of the skilled petrographers; (4) from information supplied by seismograph and gravity pendulum; (5) from any appropriate hints by astrophysicists and cosmologists; and (6) from the results of those writers who used well regulated imagination when they delved into the depths of both space and time.

Thus the granite problem demands scholarship of dismaying breadth and thoroughness as well as vision regulated by common sense. To make the best interpretation of what is both figuratively and literally a world of facts and suggestions is not science; it is rather an art, a way to science, and subject to the personal handicaps of the artist. Anyone offering a solution should be wise enough to know that his own building is in part merely a scaffolding around the temple of truth. Life is short; art is long!

Hutton showed how wrong was Werner's guess that granite was a precipitate from a primeval ocean, a world-circling "chaotic fluid." Ever since, most geologists have favored the idea that nearly all massive granite represents former magma, a melt, which during its crystallization lost some of its initially small content of gas and vapor. No widely experienced petrologist doubts that some granite-like rocks are products of metasomatism of sediments and schists, that is, the products of mineralogical change wrought by the introduction of thin, exotic, highly mobile fluids ("emanations") containing the elements of quartz and feldspar and imbibed by the invaded rocks. For a century the ratio of volume for such metasomatic granite to magmatic granite in the earth was steadily regarded as small. But during the last twenty years geologists and petrologists of increasing number have come to prefer a reversed ratio. Their doctrine of molecule-by-molecule or atom-by-atom deposition of matter essential to granite by "emanations" originating at depth, and in quality sufficient to have made most of the quartz-rich massive rocks of the world, is a kind of inverted Wernerism.

However, upholders of the new doctrine believe it possible that the energized "thin" fluids have here and there fluxed originally solid rock, with the development of true magma which has been capable of eruption in the form of bodies of "young" granite as well as bodies of rhyolite and obsidian. These exponents also recognize that partial fluxing of the kind may have made the invaded formations mobile and thus able to intrude higher levels of the earth's crust; and they also assume that during orogeny the lower part of the planet's Sial or "granitic layer" may, because of its forming part of the depressed mountain-root, have been thoroughly melted and eruptible.

Thus, while they differ about the ratio of volume of metasomatic granite to volume of magmatic granite, the two schools of thought have some common ground. This fact suggests the possible value of a convention adopted for brevity of reference and perhaps for clearer understanding of the text to follow. The convention: those who stress solidification of melts will be called "magmatists," and those who stress molecular replacement and addition of material will be called "metasomatists" The quotation marks will indicate the arbitrary character of the naming. Most writers belonging to each of the two schools emphasize the rise of matter from the depths. However, some "magmatists" prefer a two-way street, the sinking as well as the rising of material being essential in the development of granitic bodies. With one exception the "metasomatists" prefer a one-way street and the uprise of material. The "magmatists" demand temperature of 1000°C. or higher in the development, while the "metasomatists" accept much lower temperature for their emanations when still effective in the making of granite.

Another expression, also to be distinguished by quotation marks, will be convenient for present use. It is "granite" (with adjective "granitic"), intended to cover the massive quartz-rich types of rock—granodiorite, quartz monzonite, quartz syenite, tonalite, and quartz diorite, which each of the two schools explains in a manner similar to that preferred for granite proper.

The object of this paper is to list some principal reasons for the writer's belief that the "metasomatists" have overstated their case, and this because as individuals and as a group they have neglected vital facts. The grounds for his belief will appear as the discussion deals with: (1) the assumed emanations or granitizing agents; (2) the phenomena of the vast display of "granite" in the Pre-Cambrian "shields," which seem to represent fairly well the greater part of the Sial or "granitic layer" with minimum average thickness of ten kilometers; (3) the nature of a basaltic earth-shell, still deeper, which must be in close genetic connection with the formation of the Sial itself; (4) the associated questions arising from the facts known about the composition and volume of the ocean; and (5) alternative method or methods by which any large body of "granite" won space for itself in the earth's crust—a question which the "metasomatist" Read (1948, p. 7) has described as "at the heart of the granite problem," and a question particularly intriguing to any geologist who studies post-Cambrian batholiths and stocks in the field.

MODE OF ATTACK ON THE GRANITE PROBLEM

To estimate the volume-ratio of magmatic "granite" to metasomatic "granite" is necessarily a matter of theory. A good theory must rest on facts derived from observation at the earth's surface and just as obviously on assumptions about conditions in depth when "granites" were made.

At the outset it is expedient to weigh some fundamental considerations. The first of these is dry land of continental dimensions, dry land dominantly surfaced by "granite" or by the chemically equivalent acid gneisses or by the debris of "granite" and gneisses. A second fact is proof that the velocities of the earthquake waves are those expected if closely similar acid rock underlies the continental surfaces to average depth of at least ten kilometers. Geophysicists have established a third fact, that this interrupted earth-shell, the Sial, is in quasi-flotation on the earth's body because of the relatively low density of the patchy shell. The geologists have proved a fourth fact—that long before the Paleozoic era extensive dry-land masses of Sialic or dominantly "granitic" composition were eroded in some regions and veneered with quartzose sediments in other regions. The evidence for this comes from all of the existing continents. It seems therefore clear that more than 95 percent of the world's "granite" was formed far back in Archean time. And there is a fifth fact of the first rank: "granite" made in post-Cambrian time is chiefly batholithic and almost entirely confined to belts of post-Cambrian mountain-building, which themselves are located in the Sial, whether emerged or covered by the ocean.

The discussion of origins for both Sial and the comparatively young batholithic masses is laden with difficulties. We do not know the initial temperature and gas-content of the earth. We do not know whether the Moon is wholly or in part of terrestrial origin, though, if it was, the forceful ejection from one hemisphere might have produced a strong contrast of both material and temperature of the robbed hemisphere with the material and temperature of the opposite hemisphere. We do not know how effective was the Moon, supposed to be a daughter of the earth and initially close to the earth, in slowing down the planet's rotation and thus threatening the stability of the original crust of the planet by fracturing. We do not know with absolute certainty the heating power of radioactivity in the earth-shells.

Thus a speculator has to cut a number of Gordian knots before he can make any real progress toward explaining the Sial, the real "heart of the granite problem." With full consciousness of arbitrary procedure the writer does some cutting. Specifically he does not assume the temperature of the earth's outer shell in pre-Archean time to have been high enough for unmixing of a granitic layer from a basaltic layer —a conception founded on Fenner's (1948) idea of such immiscibility even in post-Cambrian time. Encouraged by the opinion of petrologists in general, the present writer assumes that important superheat does not now characterize, and perhaps has never characterized, the superficial earth-shell. He assumes that the heat of radioactivity is, and has been, steadily developed in the globe. Further, he is quite prepared to entertain the hypothesis that from the depths of the young earth steam and other gases emanated toward the surface at a rate higher than that ruling since Archean time. Therewith he finds one reason why the Archean terranes exhibit so much local metasomatism. On the present occasion he does not assume any particular thermal gradients for the earth-shells beneath the basaltic shell whence have come the basaltic lava-floods at intervals since early Archean time. He assumes that the crystallized Sialic material, if in average no more basic than granodiorite, is gravitationally stable on an underlying layer of liquid basalt. On the other hand, he assumes that, because of density difference, crystallized Sial is liable to founder in granite magma and in Sialic magma generally; also that crystallized basalt and even other somewhat less basic but crystallized rocks are liable to founder in liquid basalt. He assumes the sinking of early-formed crystals in and from either liquid. Finally, he assumes that sunken blocks are partially remelted, with the result that the more salic, lower-melting part of each rises toward the earth's surface, there being a tendency to form ultimately within the composite magma a top layer of granitic composition.

Although the "metasomatists" recognize the five great facts above listed, no member of the group has given the five all the attention they deserve in our problem. Largely because of this limited outlook the arguments for the wholesale molecular replacements have been based on widely different assumptions. The wide range of differing and competing premises essential to their scheme, is of itself ground for doubting the logic of any or all of their school of thought. Their ideas about source, chemical nature, and mode of travel of the emanations will first be examined. Then the origins of the Sial and of the post-Cambrian batholiths will be considered, when it will be found that the "metasomatists" are vague also about their second necessary premise, namely, the original character of the rock matter supposed to be molecularly replaced. Further evidence that those writers have not thought through our problem appears when the history of the basaltic earth-shell and also the composition and volume of the ocean are incidentally discussed. Thus again and again we shall see the "granite" problem is planetary and not to be solved by ordinary geological field-work.

THE "EMANATIONS"

Before taking a closer look at the arguments of the "metasomatists" it is well to note that Sederholm and his followers as well as Michel Lévy and Lacroix, all of whom by their combined researches demonstrated granitization on a comparatively large scale, are not to be classed among the "metasomatists" as above defined. For all of those Fennoscandian and French masters the source of the granitizing solutions was granitic magma. All of them were quite clear about the relative volumes of granite magma and corresponding granitized rock—clear as to which was the hen and which the egg. Equally obvious to them was the illogicality of seeking the origin of "granite" in the molecular transformation of shale, phyllite, mica schist, and quartzite, since these are chiefly detrital from "granite." All of these experts in petrogenesis were incapable of reasoning in a circle. The "metasomatists" of the present day are, therefore, not entitled to claim from these Fennoscandian and French studies any support for the idea that metasomatic "granite" dominates in the earth's crust. We may now return to the logical difficulties in the way of accepting the extreme ideas of the new school.

(1) Divergence of views about the source of the supposedly potent emanations is clear. Backlund (1936, p. 343) and MacGregor and Wilson (1939, p. 201) find a source in the solid rocks adjacent to the transformed body. Reynolds (1947, p. 222) finds there the principal source. MacGregor and Wilson (1939, p. 201), like Barth (1936, p. 830), find another source in the earth's granitic layer, the Sial. Barth (1939, p. 111) finds a source in the liquid residual from the advanced crystallization of a basaltic layer beneath the granitic Sial. Rittman (1939, p. 603) specified as source an olivine-basalt layer of great thickness and resting on a much denser planetary core. Holmes (1948, p. 414), Wegmann (1935, p. 345), van Bemmelen (1940, p. 33), and King (1947, p. 37) find the principal source at "great depth," but do not declare the nature of the parent earth-shell or earth-shells. Termier (1912, p. 592), who had recommended that the deep interior of the globe be left "aux poètes," has since written his eloquent prose to advocate colossal metasomatism and granitization by his "colonnes filtrantes" coming from great depth.

(2) Sederholm and Michel Lévy used field evidence rather than psychological necessity when they derived their granitizing solutions that bore silica and the oxides of both plagioclase and orthoclase dissolved in enough water to make the solutions "thin." Many of the more thorough-going "meta-

somatists" of the last twenty years describe the solutions as containing the oxides "needed" or "wanted" to give solid "granite" by molecular transformation of solid rocks. Since the latter include such differing types as pelites, mica schists, amphibolite, greenstone, diabase, gabbro, and limestone, the named oxides in the assumed solutions are those "needed" or "wanted" to make the "granites" whose development is being sought. Soda, silica, and lime are the oxides most commonly named, but seldom potash, which is nevertheless so important, so "needed," in the supposed granitization of limestone, greenstone, amphibolite, and basaltic rocks. In general the potash should be at least as abundant in the solutions as soda, and yet it is still to be shown why any emanation, especially one from great depth, would carry comparable proportions of the two alkalies.

Thus it is not easy to picture what the "metasomatists" mean when they write of the chemistry of the emanating solutions. Until there is clearer statement of the composition of these in terms of all the oxides and elements present, their argument for the extreme metasomatism must be regarded as weak in its second fundamental premise. Meantime specific mention of the oxides "needed" or "wanted" for granite-making might be readily taken as the product of reasoning in a circle so far as the chemistry of the postulated emanations is concerned.

One's mental trouble is not assuaged by Rittmann's (1947, p. 11 and 14) description of the granitizing agent as "gas," without his giving the composition of this gas. Nor is there any clear statement of the total composition of the rising "clouds" of ions or atoms, which, according to some "metasomatists" writing in the last dozen years, are to be preferred as the transformers. The nearest approach to such a statement appears to be that of Barth (1947, p. 181), who, though not to be classed as a full-fledged "metasomatist" as above defined, has discounted Sederholm's theory of transformism and now explains large bodies of Pre-Cambrian granite in Fennoscandia by the atomic-cloud hypothesis. For Barth each rising "cloud" is an "ichor," this term here meaning something quite different from that assigned by Sederholm, its inventor, Barth holds that his own "ichor" has produced two kinds of granite, one represented by molecularly replaced older rocks and the other represented by direct crystallizations of the atomic-cloud itself. This "ichor" is supposed by him to come from great depth. But we note again that the geophysicists are unanimously insistent on the comparatively minute proportion of potash in the deep earth-shells, and that no process of increasing the proportion of the oxide in rising emanations has ever been announced by the "metasomatists." Their dilemma is all the more apparant when we take into account a principle stated by Lapadu-Hargues (1945, p. 299), one of them, namely that the mobility, migratory power, of the atoms is proportionate to the diameter of the ions concerned, so that Na and Ca, already at depth in decided excess over K, should outstrip K in the rise toward the earth's surface. In striking contrast is the view of King (1947, p. 56) that potash travels farther than soda for example and is consequently "fixed at greater distances from source than the soda emanations."

(3) To transform a heterogeneous mass of quartzose sediments, limestone, and basic rocks into homogeneous "granite" by molecular replacement demands the expenditure of much energy, particularly thermal energy. Since, according to the "metasomatists," the visible parts of the young batholiths of the orogenic belts were so developed close to the earth's rocky surface, the granitizing agent must have brought up most of the heat required for the postulated chemical reactions. Few estimates of the agent's temperature when rising to the appropriate levels have been published. That inferred by Ramberg (1946, p. 29), when in his view Pre-Cambrian limestone was granitized, was no higher than 350°C. From other Pre-Cambrian cases Backlund (1936, p. 343) assigned "moderate" temperature, "far below the melting temperature" of ordinary rock. From Norwegian cases Barth (1929, p. 122) deduced a temperature under 500°C. for the granitizing solutions. Holmes and Reynolds (1947, p. 61), preferring an agent composed of "clouds" of ions, inferred quite moderate temperature for the reason that "ionization attains a maximum at a little above 200°C." and "at higher temperatures it begins to decrease again."

None of the four estimates is accompanied by a statement as to the depth at which the molecular transformation was effected. Presumably the emanations were much hotter before they began to rise from "great depth." On that assumption and because the rocks invaded at "great depths" were also hotter, some "metasomatists" like to think of the granitizing reactions as having taken place after the invaded formations had been orogenically depressed to depths much greater than those ruling in the uppermost part of the earth's crust.

Clearly, the "metasomatists" are not agreed about the third premise basal to their logic—that of securing sufficient energy for the making of large masses of "granite" according to their recipe.

(4) Nor are they of one mind about the mode of upward diffusion of the emanations. One of their groups favors upward migration along the mutual contacts of the minerals composing the invaded rocks. A second group favors solid diffusion, the emanational matter being ionic and rising through the space lattices of the crystals in the invaded terrane.

That the first group has been on the wrong track in its thinking is shown by Holmes and Reynolds (1947, p. 50ff), who themselves are "metasomatists" and belong to the second group. They point out that "the essential functions of gaseous or liquid solutions at any particular spot are to bring up the incoming ions and maintain their concentration, and to carry away the displaced ions. As between liquid or hydrothermal solutions and gaseous or pneumatolytic solutions, it seems probable that these functions could be carried out more effectively by the former."

Continuing, their summary statement reads: "A gas phase would have the advantage of easy penetration along submicroscopic cracks through otherwise impermeable rock, but against this it suffers from several disadvantages. Considerations of relative volatility suggest that a gas phase is most likely to be acid. In the replacement phenomena under discussion, however, the evidence is overwhelmingly suggestive of the dominance of alkaline materials. Moreover, components of widely different volatility would need to be combined in the migrating stream . . . Ingerson and Morey cite experimental observations indicating that diffusion through gas is quantitatively a more effective process than transport by gas . . . . There is the third difficulty that the degree of ionisation of a concentrated gaseous solution under high pressure and therefore of relatively high density is likely to be extremely low.

A gas phase is therefore thought to be a feeble carrier of the ions necessary to initiate solid diffusion through the crystals with which it comes into contact. Crystal linings along the surfaces of rocks and drusy cavities (rather than metasomatic replacements) are the characteristic results to be expected from gaseous transport and deposition.

"On the other hand, hydrothermal solutions, if of magmatic ancestry, are likely to be strongly alkaline from the start, as a result of hydrolysis of silicate minerals. They are also likely to have a higher degree of ionisation . . . . In respect of both composition and degree of ionisation hydrothermal solutions appear to have a definite advantage over gaseous solutions, especially in their capacity for initiating and maintaining metasomatic replacements of appropriate parageneses. But except where adequate passage-ways are available, e.g. by way of open cracks and fissures, their capacity for penetrating rocks is severely limited. Silica-bearing solutions in particular seem to have a very small power of soaking through rocks, like quartzite, in which the only passage-ways are the minute interstices between closely fitting mineral grains."

These authors therefore conclude that, since the interstices are isolated and discontinuous, "molar transport becomes impossible, and replacement can then be effected by means of solid diffusion.

"Applying the above and other relevant considerations . . . . it can be said that the few criteria available for discrimination suggest that solid diffusion was probably the dominant process concerned."

"It is impossible in a limited space to discuss the matter systematically, but to us it seems manifest that the whole process is more elegantly explained by progressive solid diffusion due to migration of ions through the crystal lattices and along the crystal boundaries."

Other writers regarding solid diffusion as the principal mechanism in granitization include Backlund (1936, p. 343), Ramberg (1948), Perrin and Roubault (1939, p. 64), Bugge (1945, p. 46), and Barth (1947, p. 181). All have rejected "ichors" as solutions. All are in duty bound to show how "clouds" of ions can make nearly or quite homogeneous "granite" in masses kilometers in thickness by diffusions into and through the crystals constituting a highly heterogeneous terrane. To discharge that duty they have assumed an enormously energized condition of the "clouds" as they start up their long upward journey, but they are far from having explained how enough of this initial energy could be retained to permit the transformism at such high levels as those near the roofs of batholiths. Neither have the diffusionists paid adequate attention to the loss of chemical potential as each "cloud" works its arduous way upward. With each centimeter of advance the "cloud" must rapidly lose its chemical potential also. Moreover, these writers have overlooked the fact that the time during which a Cordilleran batholith was formed was finite; one might well ask: which would win the race, transformism or the inevitable cooling with loss of transforming power? Is there a clear answer to that question?

A more authoritative, though in principle similar, criticism of reliance on solid diffusion in the "granite" problem was voiced at the Ottawa symposium by Bowen (1948, p. 81).

Since the day-long discussion at Ottawa saw its way into print, Ramberg (1948), a solid diffusionist, has tried to account for the upward migration of the granite-making ions. He assumes the active force to be gravity, which has caused the light, granite-making ions to rise and the "unwanted" heavier ions to sink—a case of separation and double migration. To the present writer this reliance on tractions due to the differing weights of almost infinitesmal particles inside each of a multitude of crystals is a reductio ad absurdum in the theory of solid diffusion.

To conclude: while every experienced geologist cannot doubt the rise of emanations and also the fact that rocks and minerals including ores have been modified as well as replaced by solid diffusion, neither of these actions can with good reason be thought responsible for the emplacement of the huge batholiths. In this whole matter of emanations the "metasomatists" should try to reconcile premises and lines of reasoning and, above all, to think to scale! That they have overshot the mark is still more evident as we leave the field of physical chemistry and proceed to review facts which geologists have brought home in their note-books: facts concerning the Sial as well as the "granites" of the post-Cambrian chains of mountains.

THE SIAL OR "GRANITIC LAYER"

The following brief discussion of the outer earth-shells will be preceded by a few words devoted to some questions about the interior of our planet, particularly with regard to the possibility of a superheated, vitreous state for a large part of the silicate mantle on the iron core.

Some cosmogonists are inclined to look on meteorites as giving a clue to the composition of the earth, to regard this a daughter of the Sun, and to assume an initial temperature for the planet to have been so high as to compel a molten state from surface to center. These authorities recognize immiscibility of metallic, nickel-bearing iron with the silicate material and an early gravitational separation of an iron core from its thick silicate mantle. The rate of such settling-out of the iron must have been progressively lowered as the depth increased. The enormous increase of viscosity with increasing pressure may conceivably have prevented complete separation of small droplets of iron at great depth, even after the lapse of three billion years. Moreover, there is no certainty that such high pressure did not affect the degree of immiscibility. Hence, for those reasons alone, it seems possible that the mantle never reached chemical homogeneity, thus permitting convection currents from top to bottom of the mantle. If this should have been the case, a further possibility would have to be faced: some superheat at depth. On the other hand, at and near the surface, where viscosity was low, the silicate mantile could have been thoroughly cleansed of metallic nickel-iron. There, to the depth of perhaps one or two hundred kilometers, the material can be thought to have speedily attained the composition of the iron-free meteorites, that is, a peridotite. On this assumption may be based a speculative origin for a world-circling Sial.

Jeffreys (1929, p. 138) has pictured for the infant earth a complete settling-out of the metallic iron, leaving a peridotitic mantle about 3000 kilometers thick and so homogeneous as to permit complete convective overturn of its matter, with rapid chilling, and consequent freezing of the whole thick shell. By postulating such crystallinity he considered that he could best explain the transmissibility of the shear (rigidity-controlled) seismic waves throughout the mantle. But Bridgman's (1931,

p. 340 and 1949, p. 117) experiments on the remarkable effect of high pressure on the viscosity of both liquids and gases suggest the possibility that much, if not most, of the earth's mantle may be so viscous as to behave like a crystalline solid with respect to the deforming forces connected with either earthquakes or the variation of latitude.

In his own thinking about the early differentiation of the material of the infant, star-born earth, the present writer finds it easier to assume an initially molten state for the superficial peridotite. At the temperature of about 1400°C. radiative chilling would have been extremely rapid, so that before crystal-fractionation could get under way the surface material might have been crusted to small depth—tens of meters. Following Thomson's (1883, p. 483) speculation of long ago, one should expect instability, breakup and foundering of the new crust in the thick, less dense magma beneath; some remelting of the fragments in the hotter liquid; the rise of the lower-melting, less femic material, renewed crusting and foundering, with a tendency to develop a temporarily-liquid basalt at the surface. The same result would be expected if the high-melting olivine and bronzite or hypersthene, as free individualizations, sank because of their relatively high density. By integration of both effects one can conceive that a world-circling layer of molten basalt, several tens of kilometers thick, was developed.

The molten basalt would, however, be unstable chemically. It, too, would be affected by crystal-fractionation. With a composition possibly a little more salic than the plateau-basalt of post-Cambrian eruptivity, a five-kilometer or six-kilometer, world-circling, superficial layer of material chemically close to granodiorite might result. Such differentiation can be most readily understood if it too was a step-by-step process, involving repeated crustings, repeated founderings, and repeated remelting of sunken crust-blocks. The ultimate product would thus be an unsinkable "granitic" crust resting on a layer of slowly cooling and crystallizing basalt in which the intrinsic density increased with depth so as to ensure the inner stability of this sub-layer also.

At the Ottawa symposium Bowen (1948, p. 82), leading "magmatist," did not mention the Sial or the earth's "granitic layer" and stated that the batholiths "represent the great bulk of our granites." Read (1948, p. 4), leading "metasomatist," spoke of an original granitic layer and again of "a light granitic sialic layer" overlying a denser, basaltic, Simatic layer in the earth, both layers having been formed when the planet was new. He gives no opinion as to how much of the less dense shell is true granite and how much consists of other kinds of "granite." Both of the leaders approve of the conclusion of the geophysicists who have found the upper Sialic part of the earth's crust in continental areas, to an average depth of 10 or 12 kilometers, to be "granitic." Read (1948, p. 5) seems to agree that this top sublayer was, in pre-Archean time, magmatic though even then of granitic composition. Is he, then, logical when he claims that most of the world's granite is metasomatic and that the problem of its development is a matter to be dealt with by field geology? Can the unaided field geologist deal adequately with an earth-shell so largely invisible, inaccessible?

Perhaps some "metasomatists" prefer to think of the young earth's crust as having been made of crystallized basalt or even more basic material and in pre-Archean time granitized by emanations from depth in volume to match the volume of the existing Sial. No one of their school, however, has developed the hypothesis so far as to imagine the fate of the enormous volume of material corresponding to the "basic fronts," or has deduced the effect on the volume and composition of the ocean. Any "thin" solution, representing a postulated emanation would presumably have been made "thin" by abundant water. In that case the extra silica and potash required for the granitization would be, by weight, only a small fraction of the hydrous solution, which itself was a small fraction of any earth-shell where the emanation originated. To develop a world-circling "granitic" layer only five or six kilometers thick would doubtless give an ocean several times as voluminous as that now existing. Also in all likelihood the salinity of this watery envelope would not be readily reconcilable with Clarke's (1916) careful accounting for the salinity of the existing ocean.

If the emanating solutions came from depth greater than the top of the peridotitic earth-shell, now apparently less than 80 kilometers below the surface, the basalt extruded during the formation of the Sial would, by the metasomatic hypothesis, differ chemically and decidedly from the basalt extruded in post-Archean time. Yet there is a high degree of uniformity for the basaltic lavas emitted ever since an early Pre-Cambrian epoch.

If the solutions came from such depth the transition from the Sialic layer to the layer just below it should be gradual, but some seismologists, like Gutenberg, believe the mutual contact to be relatively sharp.

The same failures to account for relevant and more or less assured facts mean as many grounds for rejecting the hypothesis of upward-migrating "clouds" of ions. Ramberg's speculation does avoid the observational difficulties by sinking the heavier ions, but, as already observed, gives an incredible picture of the way in which diffusion can change the chemical composition of any rock.

Other objections to the "metasomatic" doctrine are indicated by the proof of both seismologists and geodesists that the Sial is now largely restricted to one hemisphere. The cause of the restriction is one of the most horrendous mysteries of earth science, one not to be discussed in the present paper. But the fact is beyond cavil and of critical importance. Since the "metasomatic" hypothesis, expressed in any of its chameleon forms, recognizes no need for the emanations to choose some routes upward rather than others, the hypothesis is quite unable to account for the utter lack of granite or rhyolite within the Sial-free earth-sector capped by the deep Pacific ocean, which covers nearly one quarter of the globe. No fragment of either type of acid rock has been found in the countless basaltic flows, and no Sialic mass has been indubitably found in this sector by the seismologists.

A related fact: If the postulated emanations have been so potent as to develop the batholiths of the post-Cambrian orogenic belts, one would expect that in some form or other three million square kilometers or more of the Canadian Archean Shield should exhibit on its anciently-peneplained surface some signs of similar upward invasion; yet no tell-tale sign of new granitization since the late Archean has been found in this vast area.

Read (1948, p. 7) regards "the problem of room" as lying at the heart of the granite problem." He doubtless had in mind the batholiths and stocks of the post-Cambrian orogenic belts. How space for these bodies was secured is truly a vital question, to be later considered. It does not seem to be anything like so difficult to answer in the case of the granite of the Sial, if, as prompted by a reasonable guess, this layer was originally world-circling. Whatever the processes responsible for that vast granite-making, they had "all outdoors" in which to carry on their constructive work. If, on the other hand, the "granitic layer" was formed almost wholly in one hemisphere the colloquialism would not be strictly true. In either case the emplacement of the Sialic granite would not have involved the same kind of structural problem as that so insistent at post-Cambrian batholith or stock. Yet this conclusion does not conflict with the idea that molar movements were essential for the development of both Sial and batholithic bodies.

The writer also holds the opinion that the areal restriction of the Sial is not easy to reconcile with the view of a "magmatist" who believes each body of "granite" to be a direct, single-step differentiate of basaltic magma. During the long lives of the Hawaiian volcanoes the basaltic magma below their vents was chilled countless times to temperatures below which crystal-fractionation should, by theory, have produced "granite" or its effusive equivalent. Yet neither of these rock types has been discovered in the thousand-mile chain of islands. The same is true for some other volcanic piles of the central Pacific. Again, if the basaltic magma under the Canadian "shield" has been slowly freezing since late Pre-Cambrian time, with the direct formation of a corresponding amount of "granitic" differentiate, some of this relatively light material should have risen to the surface or to crust-levels reached by erosion. But there is no record of such upsqueezings since the beginning of the Paleozoic era.

Nor is it likely that the material of the Sial could represent the residual liquid resulting from the crystal-fractionation of a peridotitic earth-shell in a single step. The writer's doubt on this point is based on an application of the Clausius-Clapeyron equation connecting pressure with the temperature of freezing From this relation it appears that the granitic residuum of crystallizing peridotite would be itself pressure-crystallized before it could be squeezed out of a peridotitic layer with top 40 kilometers or more below the earth's surface the hypothesis of replacement by "granitic" magma, along a two-way street, where stoping, remelting at some depth, and renewed differentiation are assumed.

Incidentally it may be recalled that not only all "metasomatists" but also nearly all the leading "magmatists," from Rosenbusch, Osann, Zirkel, and Teall, to Lacroix, Shand, and Bowen, have disregarded magmatic stoping as an essential process in the generation of the world's "granite," whether Archean (Sialic) or post-Cambrian.

Holmes (1945, pp. 91, 409), a convinced "metasomatist," has raised several objections to the stoping hypothesis. First he notes that with increase of depth xenoliths in "granite" "become gradually smaller and less numerous, and more and more like the granitic rocks enclosing them, until finally they disappear altogether. Evidently they have been incorporated in the granite and the space problem remains." But the "magmatist" has a different answer: the viscosity of his "granitic" magma at the upper levels of the cooling mass increases at maximum speed until the enclosing material attains the small strength sufficient to prevent further sinking, while lower down the liquidity persists much longer and thus permits complete cleansing of xenoliths.

Holmes's second objection reads: "Long 'islands' of country rocks like those of the metamorphic aureole can sometimes be recognized in the granite, their structures remaining in continuity with those of the bordering country rocks, though they may often be no more than vague shadowy outlines, seen through a veil of granitization. Black metamorphosed dolerite dykes, older than the batholith, can sometimes be traced into it as partly granitized 'ghosts.' Since the heavy dyke rocks did not sink, there is no reason to suppose that the associated metamorphosed sediments could have done so." A "magmatist" answer is that the "islands" are roof-pendants, stably supported in the "granitic" melt by their own strength, which with downward tapering of each pendant can be shown to ensure such stability until the whole batholith freezes solid.

The most serious objection is the third: "If granite magma ascends in great bulk towards the surface in this way, it should often have broken clean through the crust to form gigantic volcanoes erupting rhyolite and obsidian and the corresponding pyroclasts. But this has very rarely happened, as the retention of the original roof clearly proves." This difficulty for the stoping hypothesis was fully apparent when the hypothesis was first proposed but the present writer has steadily believed the difficulty can be surmounted. A considerable number of cases are recorded where large bodies of rhyolite, dacite, and obsidian have been erupted in areas where masses of apparently syngenetic "granites" were developed. A remarkable example, recently described by P. Lapadu-Hargues (1947, p. 950), is in the Eglab massif of the western Sahara, where at intervals for 150 to 200 kilometers a huge granite batholith passes continuously upward into broad cappings of rhyolite. It is, of course, conceivable that in many other mountain belts where erosion was prolonged similar rhyolitic cappings were completely destroyed. Nevertheless, it is probable that in perhaps most instances "granitic" batholiths did not break through to the earth's surface. A leading reason for this may well have been the freezing of the magma, chilled by the stoping itself as well as by conduction of the magmatic heat into rock which the stoping magma found to be cooler and cooler as it made its way upward. Perhaps, too, progressive loss of gas and corresponding rise of the freezing temperature is to be considered. On various grounds, therefore, the writer does not regard Holmes's third objection as compelling.

His fourth reads· "If granite magma rose from the depths in quantities corresponding to the enormous volumes of batholiths, it would be by far the most abundant of all magmas. In this case rhyolite should be the most abundant of all volcanic rocks. But it is not." For the "magmatist" "granite" is actually by far the most abundant of the massive rocks in the orogenic belts, the regions now under discussion. We have just found good grounds for doubting the added remark about rhyolite in these same belts.

Holmes (1945, p. 409) is willing to entertain the idea that convection to great depth in the earth may be competent to cause mountain-building of the Cordilleran type. This action, if real, must have resulted from the exceedingly small potential due to relatively small difference of temperature in the thick earth-shell involved. Yet he is averse to crediting the reality of solid-liquid convection to much smaller depth, though the potential is here incomparably greater and the overturn of material incomparably smaller in quantity. One must conclude, therefore, that Holmes in his general review of physical geology should not have had a priori objection to stoping by acid magma to the depth of ten or twenty kilometers. Since more thoroughly than any other geologist he has expressed doubts about the efficiency of magmatic stoping, analysis of arguments against the hypothesis seems not to invalidate it as easing explanation of post-Cambrian batholiths and stocks.

For those bodies the difficulties of the "metasomatists" include some of those already listed when the origin of the Sial was considered, but some of the arguments proffered by members of the school deserve special attention. But beforehand a statement by Bowen (1948, p. 88) at the Ottawa symposium may be recalled: "Though, locally, volume for volume replacement of individual beds may be demonstrated, the claim that wet granitization offers a solution of the 'room problem' is altogether spurious. Material pushed up chemically occupies just as much room as the same material pushed up mechanically. Even though repeated equivolume replacement is postulated, it cannot be repeated *ad infinitum.* There would necessarily be a large net addition of material in the upper crust. Wet granitization does not solve, it merely evades the room problem. The magmatist has, however, found not unreasonable real solutions of the problem."

In whatever way the "metasomatists" account for the earth's "granitic layer," they must admit that, according to their own views, the bulk of each post-Cambrian batholith of "granite" was itself metasomatized Archean (Sialic) "granite" and its own derived sedimentary products. Would "emanations" develop in the post-Cambrian orogenic belts "batholithic" rock-types like those of the primitive earth-shell, which, unless all signs fail, is best represented in visible presence by the oldest Archean terranes? The fact is that, apart from the problematical charnockite and the equally mysterious gneissic structure of these oldest "granites," dominant rock types of the Archean match well with the dominant types in the post-Cambrian batholiths and stocks.

Another trouble for the "metasomatists" in their explanation of these bodies is their comparative homogeneity in the vertical direction, a characteristic often displayed also over outcrop areas measurable in hundreds of square kilometers. An example is the batholith crossed by the Similkameen River of Washington State and British Columbia (Daly, 1912, p. 435). There the mass extends from the highest point in the valley wall down to the river nearly 2000 meters lower down. No floor is to be seen anywhere and the body has lost by erosion an unknown amount of substance above its highest outcrop. Many other batholiths of Alaska, British Columbia, and California afford proof of great thickness and total volume as well as comparative homogeneity, and there can be little doubt that these regions fairly represent the grandeur of the batholithic development along the cordilleran chain stretching from northern Alaska to Cape Horn—the world's most spectacular display of post-Cambrian "granite."

The almost perfect homogeneity of many batholiths to visible depths is by the "metasomatic" doctrine hard to understand not only because of the pronounced heterogeneity of the invaded rocks but also because of the strong probability that, if the emanations were fluids, made "thin" by the presence of much water, one would expect the upper part of each botholith to be specially rich in minerals stable in the presence of much water. Yet there is no field evidence of such differential hydration in the post-Cambrian bodies.

Further, no protagonist of the doctrine has shown why those bodies are confined to orogenic belts or, as already observed, why diffusion could have accomplished the enormous molecular replacement in the limited duration of an orogenic revolution.

Those who emphasize upward diffusion of emanations from great depth cannot doubt that magmatic basalt has risen along each belt both before and after the paroxysm of mountain-building and nevertheless retained the same composition. Yet by the metasomatic hypothesis the basalt itself should have been chemically changed while deep-sourced emanations were passing through it to the high batholithic levels.

Again, the observer exploring the high cliffs of the Cordillera sees that the lateral, steeply inclined contacts of the "granitic" giants as well as the contacts of enclosed xenoliths are commonly if not generally sharp. He notes there the relative slightness of metasomatism. He discovers the never-failing fringe of apophysal dikes from batholith and stock. He may think of the old masters. Will he find more comfort in the vaguely defined "plutonism" of Lyell than in Hutton's exciting demonstration of magmatic granite?

That some metasomatism has accompanied or immediately followed the emplacement of Cordilleran batholiths is clear, but their most experienced students have no doubt as to which is the horse and which the cart. They have found that such molecular replacement is much more conspicuous at the roof of a batholith than along its lateral contacts. The most remarkable example, quantitatively considered, is that reported by G. H. Anderson (1937). After prolonged exploration he explained the "Pellisier granite" of the California Inyo Range as the product of metasomatism of Cambrian and Pre-Cambrian schists by emanations from beneath. If he is right, he has found the largest Cordillera body proved to have been developed by metasomatism. He pictures it as a roof phenomenon. The Pellisier "granite" rests on the Boundary Peak granite, which in the molten, batholithic state gave forth the replacing fluids. From his cross-section one sees that Anderson assigned the remarkable thickness of 600 meters to 900 meters to the "Pellisier granite" while the parent Boundary Peak granite is two or three times thicker, even though no bottom for it appears in the draughted section. He too had no doubt as to which was the hen and which the egg. Like Sederholm and Michel Lévy he felt no need of seeking in mystical depths the immediate source of the working fluid.

CONCLUSION

To summarize: it seems evident that "metasomatists" have left gaps of critical importance in every published version of their hypothesis. They should make more specific: (1) their ideas concerning the nature of the Sial and its mode and time of development, (2) their premises with respect to the state, composition, source, energy, and mode of travel of the assumed emanations; (3) their estimates of the volumes of magmatic "granite" formed by melting of mountain roots, and their estimates of the volumes of "granite" made molten by anatexis; (4) their explanation of the fact that during at least half a billion years no discernible granitization of the earth's crust has been registered under the vast, deep Pacific or within nearly all of the extensive Canadian "shield"; (5) the rôle of crystal-fractionation plus remelt-fractionation in petrogenesis; (6) truly compelling grounds for doubting large-scale magmatic stoping, whether in relation to the earth's "granitic layer" as a whole or in relation to batholiths and stocks of the mountain chains; (7) explanation of the absence of material in the visible part of the earth's crust, corresponding to the "basic fronts" expected by the hypothesis, (8) the meaning of the almost perfect chemical uniformity of basaltic magma erupted since a comparatively early stage of the Archean; (9) the volume and composition of the ocean in relation to the hypothesis; (10) the commonly displayed sharpness of contact of post-Cambrian batholith or stock with the respective invaded formations; (11) the nearly perfect chemical and mineralogical uniformity of so many of these subjacent masses in the vertical direction; and (12) the restriction of typical post-Cambrian bodies of the sort to the mountain chains.

Until all the twelve kinds of argumentative trouble are overcome the "metasomatists" cannot be said to have succeeded in their effort to account for most of the world's "granite". As briefly indicated in the body of the foregoing text, there seems to be a more excellent way, though it too has to be speculative and reach back into the mists of cosmogonic theory.

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UNUSUAL STYLOLITES*

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ABSTRACT. Stylolite seams occurring at near right angles to bedding planes are found in certain limestones along the eastern flank of the Front Range and Wet Mountains of Colorado, where known thrust faults exist

Limestones of the Pennsylvanian Fountain, Jurassic Morrison, and Cretaceous Greenhorn and Niobrara (Timpas) formations were studied in detail The stylolite seams intersect bedding planes at remarkably constant angles whether they occur in undeformed, normally dipping, or strongly over-turned strata. The seams are only roughly parallel to each other, but the axes of the columns show almost no variation in attitude within an exposure

It is believed that the stylolites formed during the early stages of Laramide orogeny. A system of transverse joints developed in the relatively undisturbed beds of the section and solution along these joints, combined with continuing stress associated with the thrusting produced the stylolites The major deformation of the beds, including local over-turning, is later than the stylolites

INTRODUCTION

Because of recent discussion regarding the origin of stylolites; Goldman (1940), Shaub (1939), Stockdale (1943) and others; the nature and occurrence of these structures in certain limestones near Colorado Springs, Colorado, is believed to be significant. The evidence, besides firmly establishing the mode of origin, suggests two possible uses of stylolites in the interpretation of structural problems.

The area of detailed study extends from the Colorado Springs region southwest along the east flank of the Colorado Front Range, across the Canon City Embayment to the east flank of the Wet Mountains. The sedimentary formations on the eastern slopes of the mountains have been deformed and show dips ranging from a few degrees to vertical, with overturning at some localities along both mountain fronts. The formations dip gently to the southeast in the Canon City Embayment, but are almost horizontal in the center of the embayment at Portland, Colorado.

The stylolites are found in limestones of the Pennsylvanian Fountain formation, the Jurassic Morrison formation, and the Cretaceous Greenhorn and Niobrara (Timpas) formations. They are most numerous, best developed, and best exposed in the Timpas member of the Niobrara formation. Stylolites of each formation will be discussed separately.

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MODE OF OCCURRENCE

PENNSYLVANIAN FOUNTAIN FORMATION

Lithology: Limestones, which comprise a very small part of the Fountain, occur in the upper one hundred feet of the formation. The limestone is gray to reddish-gray, dense, commonly laminated, and exhibits good conchoidal fracture. It occurs in discontinuous beds or lenses which are locally nodular. Irregular white crystalline calcite masses, averaging one quarter inch in diameter, commonly occur in the rock. Wavy laminae of poorly indurated red shale one-eighth inch thick are interbedded with the limestone.

The lithology of the Fountain is extremely complex and individual beds can be traced for only short distances.

Stylolites. Stylolites were observed in the vertical and slightly over-turned beds of Fountain limestone in the low saddle south of Keeton's ranch house on Little Fountain Creek.* The seams are generally parallel to each other and intersect bedding planes at angles of about eighty degrees. The individual stylolite columns are cone-shaped and were observed to have a maximum height of three quarters of an inch, the average being one quarter inch.

At a road cut in Deadman Canyon, stylolites were observed in gently dipping beds of limestone. Some seams cut the bedding planes at steep angles. Other seams, anastomosing considerably, occur parallel to the stratification, but the two sets were not observed to cross. Of the seams parallel to bedding planes, many are unquestionably stylolites, yet some of the same seams appear only as wavy shale partings when traced several inches. The thickness of the shaly material in these undulations is several millimeters, but in the stylolitic portions along the same plane it is only about two-tenths of a millimeter. Since a greater thickess of shaly material occurs in the wavy partings, there appears little possibility that the undulations are embryonic or poorly developed stylolites. Probably these lenticular shale seams are depositional, the thinner edges of the seam being secondarily developed into stylolites, whereas the thicker portions of shale prevented the interpenetration of the limestone to form stylolites.

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* Most of the specific localities mentioned in this paper may be found on the Colorado Springs Quadrangle map or on the new maps, now in press, of the Cheyenne Mountain, Big Chief Mountain, Lytle, and Mount Pittsburg Quadrangles which cover the southwestern quarter of the old quadrangle

Of the transverse stylolites observed in the Fountain, all intersect bedding planes at steep angles whether the beds be over-turned or dipping only slightly.

JURASSIC MORRISON FORMATION

Lithology: Limestone, especially gray, lithographic or sublithographic type comprises a minor part of the Morrison and occurs about sixty feet above the base of the formation. The beds are from two to twenty-four inches thick and locally there is fine lamination.

Although the lithology of the Morrison is extremely variable within the area, the limestone beds are persistent and of essentially uniform lithology.

Stylolites: Stylolites are not particularly common in the Morrison. The best locality is at a small, abandoned tunnel one hundred yards east of Colorado Highway 115 near the large road-cut at the south end of Deadman Canyon.

The stylolite seams are transverse, intersecting the bedding planes at about eighty degrees. The stylolite columns are conical and have a maximum height of three-quarters inch and an average of one-quarter inch.

CRETACEOUS GREENHORN FORMATION

Lithology: The Greenhorn consists of limestone beds from three to eighteen inches thick separated by shale beds from six to forty-eight inches thick. The limestone is a dense, gray, argillaceous rock with good conchoidal fracture. The shale is light gray, poorly indurated, and calcareous. Thin bentonite beds occur locally in the shales, and veins of gypsum, generally less than a quarter of an inch thick, are common. The lithology of the Greenhorn is uniform throughout the area and the dominance of shale in the section makes good exposures of the limestone beds rare.

Stylolites: Although only a few stylolites were observed in the Greenhorn, they are all of the transverse variety. The seams are roughly parallel and intersect bedding planes at angles of about seventy-five degrees. The nature and occurrence of known stylolites in the Greenhorn are similar in every respect to the transverse stylolites of the other formations studied.

CRETACEOUS NIOBRARA (TIMPAS) FORMATION

Lithology: The lower part of the Timpas consists of limestone beds from three to twenty-four inches thick separated by shale beds from one-half to ten inches thick. It is in this portion that the stylolites are best developed.

The limestone is a gray to bluish-gray, dense, argillaceous rock with good conchoidal fracture and containing numerous *Inoceramus deformis* shells and shell fragments. Iron sulfide concretions are common. The shales are light to medium gray, poorly indurated and calcareous. Lamination is generally present but is best developed along contacts between shale and limestone, which appear to be gradational.

The lithology of the Timpas is remarkably constant throughout the area.

Stylolites: Stylolites may be observed in almost any good exposure of the lower Timpas, but one of the best and most accessible localities is an abandoned roadside quarry on Colorado Highway 115 just south of Little Fountain Creek.

TABLE I

Attitude Relationships of Beds and Stylolite Seams

Formation & Age	Dip & Strike of Stratum	Dip & Strike of Seam	Approx Angle of Intersection
Niobrara (Timpas) U. Cret.	N14W 30N	N56E 73N	87 degrees
	N15W 24N	N50E 69N	82 "
	N25W 27N	N48E 76N	84 "
	N11W 25N	N42E 62N	80 "
	N28W 23N	N33E 47N	62 "
	N 7W 32N	N52E 68N	87 "
	N23W 34N	N29E 71N	86 "
	N22W 26N	N35E 57N	73 "
	N14W 20N	N58E 79N	86 "
	N10W 24N	N44E 57N	67 "
	N 5W 7N	N51E 80N	83 "
	N 8W 8N	N33E 64N	71 "
	N20W 11N	N41E 70N	78 "
	N 2W 11N	N23E 73N	81 "
	N12W 8N	N40E 70N	74 "
Greenhorn U. Cret	N12W 30N	N28E 52N	75 "
Morrison Jurassic	N10W 24N	N74E 80N	83 "
Fountain Penn	N50E 57N -overturned	N45E 28S	80 "

The stylolite seams are distinctly transverse to the bedding of the limestone. Except for minor irregularities and branching, the seams are roughly parallel to each other in the exposure. The planes of the seams intersect the bedding planes at angles averaging about eighty degrees (Table I, figs. 1, 2).

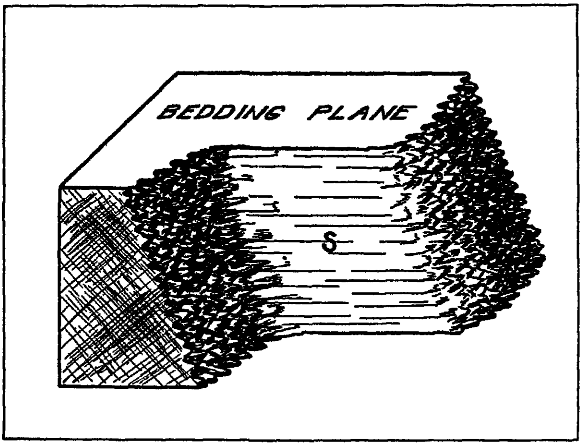

Figure 1 Stylolite seam offset along slickensided plane (s) which is parallel to cone-axes of the stylolite

Almost all of the stylolites are of the angular type with individual penetrating elements being cone-shaped. The individual cones are not necessarily oriented normal to the plane of the seam, but may be inclined at any angle up to ninety degrees, depending upon variations in the strike of the seam. In any case, the axes of the cones are parallel throughout any exposure. Some of the seams were noted gradually to change strike almost ninety degrees within a distance of several feet, and then similarly to change back to the original direction several feet farther away. The stylolite cones are inclined steeply where the seam trends in the usual direction, but where the seam gradually changes strike, the cones meet the seam at shallower angles until, at the greatest deviation of the seam, the cones lose their identity and ordinary slickensides appear. Crystalline calcite surfaces one-quarter inch thick comprise the slickensided portions of the seam. Where the normal trend of the seam is resumed, stylolites once again appear. The slickensides have an alignment identical to the attitude of the cone axes (fig. 1).

The observed height of the columns varies from minute to a maximum of four inches, the average height being about a quarter of an inch. The largest stylolites are of the columnar type, having columns about two inches wide. The extremities of these stylolites consist of numerous small cones identical to those forming the smaller stylolites. Perhaps the columnar

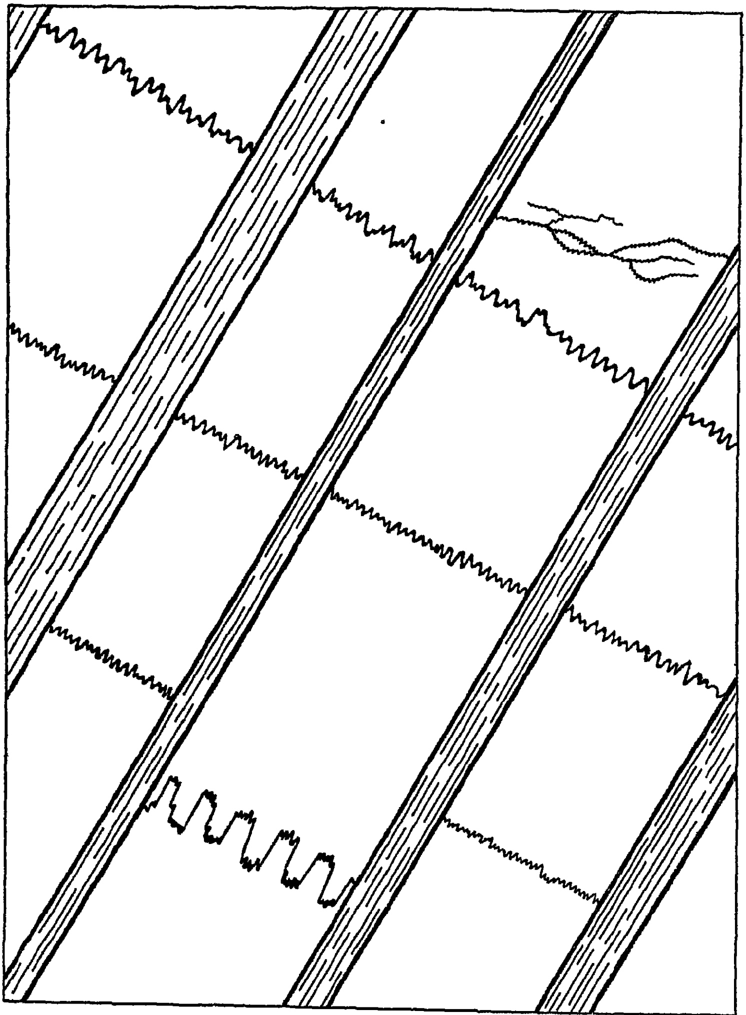

Figure 2 Composite diagram showing typical relationships of stylolite seams to limestone beds, white; and shale beds, shaded; in Niobrara (Timpas) formation

type is a characteristic of only the larger, more completely developed stylolites, the conical type being limited to the smaller ones.

The transverse seams may or may not be restricted to a single limestone stratum. Stylolites with low amplitudes were observed to disappear when traced into ten-inch shale beds, only to reappear in numerous succeeding limestone strata, the attitude of the seams remaining constant throughout. However, some of the stylolites having greatest amplitudes are restricted to a single limestone stratum with no trace of the seam appearing in adjacent shales or limestones, although the shale parting may be a fraction of an inch thick. Therefore the persistence of a seam bears no relationship to the size of the stylolites or to the thickness of the shale beds which separate the limestone strata.

Stylolites were observed parallel to bedding planes, but the occurrence of these is rare. This type is found in limestone beds of the lower Timpas only within about two inches of the bedding planes, where lamination is common. Transverse stylolites were noted to cross the other set. At the intersection, there appears a discordance on the parallel type seam which indicates the transverse seams to be younger.

A false impression of the seam attitudes results if the stylolites are observed only in cross section. The true seam attitude can be determined best by observing the weathered surfaces of the limestone where the planes of the seams are exposed completely. The erroneous impression of attitude is well exemplified by the larger stylolites having amplitudes of several inches. If the limestone is fractured across the seam at a very low angle, haphazard stylolites appear in cross section because of the decapitation of the columns. This also gives the false impression that a number of erratic seams are present, whereas only one large, rather uniform seam occurs, which is related in attitude to the other true seams.

AREAL DISTRIBUTION

Within the area of study, stylolites are most numerous and best developed along the flanks of the Front Range and Wet Mountains, especially where known thrust faults exist. The Cheyenne Mountain thrust marks the steep eastern face of the Front Range from Colorado Springs to the vicinity of

Little Fountain Creek. In most places where the Timpas crops out within these limits, transverse stylolites are very abundant. Beyond the mappable trace of the fault to the southwest, and at exposures in the Canon City Embayment, only a few transverse stylolites were observed, these having amplitudes of less than one-eighth inch. Southwest of Canon City, Colorado, on the west side of the embayment, the sedimentary beds are deformed by the Wet Mountain thrust, and show normal dips to the east, but are over-turned locally. Numerous well developed transverse stylolites were observed in limestone beds dipping about forty degrees to the east. The relationships of the seams and beds are comparable to those already discussed.

ORIGIN

Beyond doubt, these structures are stylolites in spite of the transverse mode of occurrence By studying a hand specimen containing the structures, any observer would identify them as stylolites. The transverse attitude, therefore, cannot be used as a criterion to eliminate them as stylolites.

The main principles of Shaub's "contraction-pressure" theory of primary origin of stylolites will be briefly discussed. Shaub (1939) states that impermeable clay seams trap upward moving pore water in limestone oozes underlying the clay. The upper part of the ooze becomes more plastic than the lower part as water accumulates. A slow drying of the ooze results in stress differences and places of low pressure. The ooze, under the force of gravity or other pressures, flows into the low pressure areas shearing the clay band, the result being a stylolite seam.

Shaub's theory provides no adequate explanation for stylolites transverse to normal bedding. The occurrence of this type, on the other hand, contradicts Shaub's main principles. Clay partings along transverse stylolites could not have been originally deposited as clay seams. Even if this were possible, upward moving pore water would not accumulate under clay seams dipping eighty degrees.

Furthermore, the occurrence of transverse stylolites with remarkable constant characteristics in formations ranging in age from Pennsylvanian to Upper Cretaceous precludes any possibility of a primary origin. Because stylolite seams of all formations studied intersect bedding planes at angles remarkably alike, it is inconceivable that the stylolites of each formation could have formed successively as each formation was deposited.

The most reasonable conclusion is that after the deposition of the Cretaceous beds, the relatively undisturbed beds were jointed by the early movements of the Laramide Revolution. Solution activity along the transverse joints, under the influence of continued stress associated with thrusting, caused the development of stylolites simultaneously in the limestones of various formations. Beyond the influence of thrusting, especially in the Canon City Embayment where the beds are essentially horizontal, stylolites are lacking. Following the stylolite development, the major thrusting deformed the strata and caused local over-turning.

The largest and best developed stylolites usually conform to the general alignment, but the smaller ones are irregular, commonly branching and anastomosing. The former correspond to the best original joint development, the latter to smaller and more irregular joints and fractures, which are common in dense limestones.

The stylolite columns are always oriented perpendicular to the direction of pressure, which must have been constant over wide areas. Because of irregularities in the original joints, individual seams have varying strikes. Therefore the columns, although mutually parallel within an exposure, meet the seams at varying angles of inclination. Where a stylolite seam is inclined to the direction of pressure, the adjacent limestone blocks could not slide past each other, but rather had to interpenetrate as the limestone was removed by solution along the joints. But where the joint is essentially parallel to the direction of pressure, the adjacent blocks easily slid past each other forming slickensides.

It is generally agreed that slickensides may be produced by small displacements of the rocks along a fracture. The slickensides on fractures parallel to the stylolite columns were produced by movement not exceeding that involved in the making of the stylolite. With stylolites averaging less than half an inch in amplitude and having very thin residual clay seams the movement was certainly small yet excellent slickensides were developed.

The association of stylolites and slickensides proves beyond doubt the importance of pressure in the formation of stylolites, since the slickensides are genetically dependent upon pressures to cause relative movement of the adjacent blocks. The results of Stockdale's work (1943) show that solutions dissolve rock material most easily at points of greatest pressure. The greatest pressure along a joint would be in the portion which is most nearly perpendicular to the pressure direction, and the points of lowest pressure would be where the joints are parallel to the pressure direction. Therefore it is conceivable that calcium carbonate was dissolved at the higher pressure portion of the original joints where stylolites formed, and that it was redeposited in the low pressure portion, being slickensided by the relative movement of the blocks, or inheriting the slickensides from the adjacent limestones.

The persistence of a stylolite seam depends on the extent of the original joint. The angle at which the stylolite seams intersect the bedding planes is that of a system of joints common to all the formations studied within the area.

It is evident that the only theory providing adequate principles regarding the origin of these transverse stylolites is the "solution-pressure" theory of secondary origin of stylolites as presented by Stockdale. In this area the pressure was orogenic rather than static as is the case in better known areas of stylolites.

PROPOSED USES

The study of transverse stylolites may be of structural significance in two ways:

1. To determine the direction of regional stresses.
2. To determine top and bottom of beds and therefore the structural attitude of isolated outcrops.

Direction of regional stresses: Because most, if not all, stylolites are developed by combined solution and pressure the cone axes will be parallel to the pressure direction. For transverse stylolite seams the cone axes will indicate the direction of regional stresses whereas stylolite seams parallel to the bedding indicate the stress of load.

Determination of top and bottom of beds: The present opinion regarding the significance of stylolites in determining top and bottom is best stated by Shrock (1948, p. 240):

"There seems to have been no mention of any characteristic of stylolites which might make them useful in determining the top or bottom of a sequence in which they occur."

Under circumstances similar to those in the area studied it may be possible to use stylolites as a key to structural relationships. Here it is possible to study the stylolites in numerous exposures where top and bottom of beds and structural attitude can be determined with certainty by various criteria. These studies show that:

1. The attitude of the stylolite seams with respect to the bedding planes is approximately constant without regard to age of the limestones or structural attitude.
2. The attitude of the stylolite cones has an even more constant relationship to the plane of the bedding than does the attitude of the stylolite seams, and that the cones are inclined rather than normal to the seam.
3. The stylolites developed by solution and pressure along joints in the limestone before the major deformation, including local over-turning, occurred.
4. The strike of the stylolite seams is appreciably different than the strike of the beds.

If, in such an area, there are isolated exposures of stylolitic limestones in which top and bottom of the beds and therefore the true structural attitude cannot be determined by other criteria the stylolites may provide an adequate criterion.

The limiting directions and angles of plunge of the stylolite cones can be determined by means of a simple model consisting

TABLE II

Attitude Relationships of Cone Axes in Beds Known To Be Dipping Normally

Formation	Attitude of Bed	Directions and Plunge of Cone Axes
Niobrara (Timpas)	N 8W 26N	S69E 16
	N 6W 26N	S64E 17
	N 5W 30N	S66E 22
	N10W 28N	S62E 21
	N 4W 29N	S66E 21
	N 3W 27N	S63E 19
	N 6W 29N	S66E 20
	N 2W 30N	S63E 16
	N16W 10N	S65E 10
	N15W 8N	S70E 10
Morrison	N10W 24N	S70E 20

TABLE III

Attitude Relationships of Cone Axes in Beds Known To Be Overturned

Formation	Attitude of Bed	Directions and Plunge of Cone Axes
Niobrara (Timpas)	N 8W 54S N 2E 59N N10W 47S N20E 45N	S25W 25 S25W 30 S33W 30 S20W 25
Fountain	N30E 75N	S40W 5

of a cardboard box and a pencil. The box is used to represent the bed and is placed in measured attitude. The pencil is inserted at the measured direction and plunge of cone axes. The box is then rotated normal to its strike and the possible directions of plunge of the stylolite cones are noted for cone axes in beds not over-turned. Continued rotation of the box will demonstrate the cone attitudes in over-turned beds. Because the cone axes are nearly parallel to bedding planes, the direction of plunge will change at the first over-turning. Thus, over-turned beds could be recognized by a plunge direction of cone axes which could not possibly exist in beds of a normal depositional sequence (Tables II and III).

ACKNOWLEDGMENTS

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SPHEROIDAL STRUCTURES IN ARIZONA VOLCANICS

JOHN H. FETH AND JOHN W. ANTHONY

ABSTRACT. Spheroidal forms occurring in a red welded keratophyre tuff and in a devitrified gray tuff in the Canelo Hills, southeastern Arizona, are described

The keratophyre tuff has been traced for approximately one mile. Thickness ranges from 20 to 25 feet. The spheroids range from a fraction of an inch to 12 inches in diameter, and constitute from 50 to 75 percent of the weathered surface throughout its entire extent. Flow structure, apparently continuous through both matrix and spheroids, is clearly displayed

The spheroids in the gray tuff have been found in an area measuring about 600 by 200 yards, and occur about 100 feet lower in the volcanic sequence than the welded tuff They are less abundant and less prominent on the outcrop than the spheroids in the welded keratophyre tuff, but occur in many places free from matrix in the talus. All have cores, most of which are tesselated calcite.

Thin sections were prepared to encompass entire equatorial cross-sections of individual spheroids

Under the microscope sporadic spherulites were found within the larger spheroids in the welded tuff Well developed flow structure is characteristic. Both primary and secondary quartz are abundant. An indication of radial orientation of quartz and feldspar grains suggests that the growth of the spheroids was determined by a primary control Absence of demonstrable secondary control points toward the same conclusion

The gray tuff shows heavy concentrations of secondary calcite other than that in the cores, impartially distributed in matrix and spheroids. Comparable structures in volcanic rocks from other Arizona localities are mentioned.

THE occurrence of spheroidal structures in profusion in a red keratophyre tuff and the presence of comparable nodules in an adjacent gray tuff engaged the attention of the senior author during field mapping in the northern Canelo Hills, Santa Cruz County, Arizona. The Canelo Hills lie sixty road miles southeast of Tucson and twenty-five miles east and north of Nogales. They form a connecting link of high land between the Huachuca Mountains to the east and the Santa Rita Mountains to the west.

The series of volcanic rocks in the area includes rhyolites, andesites, and tuffs, and has a thickness of 991 feet on Mount Hughes, two miles distant from the locality where the spheroidal structures are most plentiful. Schrader (1915, pp. 70-75) considers the volcanic series to be Tertiary in age. No more recent data concerning their age has been published as far as is known.

The volcanic rocks were extruded over an eroded surface of Cretaceous (?) sediments, and have been gently warped in the northern Canelo Hills into a broad anticline whose axis trends northwest-southeast. These volcanic rocks are covered in part by a deeply eroded block of Permian sediments, principally limestones, overthurst upon the area, and in part by a fanglomerate composed of cobbles and pebbles of Permian limestone and of the volcanic materials. The rocks in the area are cut by high-angle faults, one set trending approximately north-south and a second approximately east-west.

ACKNOWLEDGMENT

The writers acknowledge with appreciation assistance and suggestions given in the laboratory by T. S. Lovering, Felix Gonzales-Bonorino, and Hal Morris. B. S. Butler, E. D. McKee, M. N. Short, and E. D. Wilson have read the manuscript and made numerous suggestions which have been followed in the final draft.

FIELD OCCURRENCE

Spheroids occur in a dark red keratophyre tuff 20 to 25 feet thick. They range in diameter from a fraction of an inch to about 12 inches, and weather out on the surface of the tuff, to form a nodular structure of striking appearance (pl. 1, fig. 1). As much as 75 percent of the outcrop is covered by the nodules in localities where the rock stands out from a slope, as on low cliffs. Elsewhere the spheroids thickly litter hilltops with rounded masses. The outcrop extends for more than a mile without noticeable diminution in the quantity of spheroids. The outcrop is terminated at the southwest by a high-angle fault. There the tuff is in contact with the Cretaceous (?) series which is exposed in a canyon where the overthrust Permian limestones cap the opposite canyon wall. To the northeast the tuff is cut off by a high-angle fault.

It is noteworthy that flow lines in the keratophyre tuff are continuous through matrix and spheroids alike, but that the spheroids are separated from the matrix by a sharp line of demarcation. The spheroids weather out from the matrix, leaving clearly defined hemispheric cavities on the outcrop. On weathered surfaces many of the spheroids exhibit a warty character analogous to the "cauliflower-like habit" mentioned

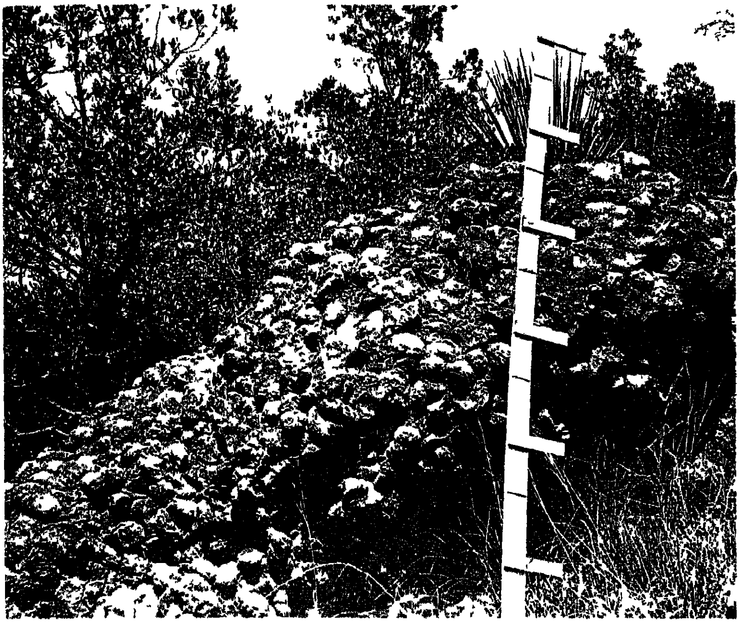

Fig. 1 Outcrop of welded keratophyre tuff showing profuse development of spheroidal structures Horizontal projections on staff 12 inches apart

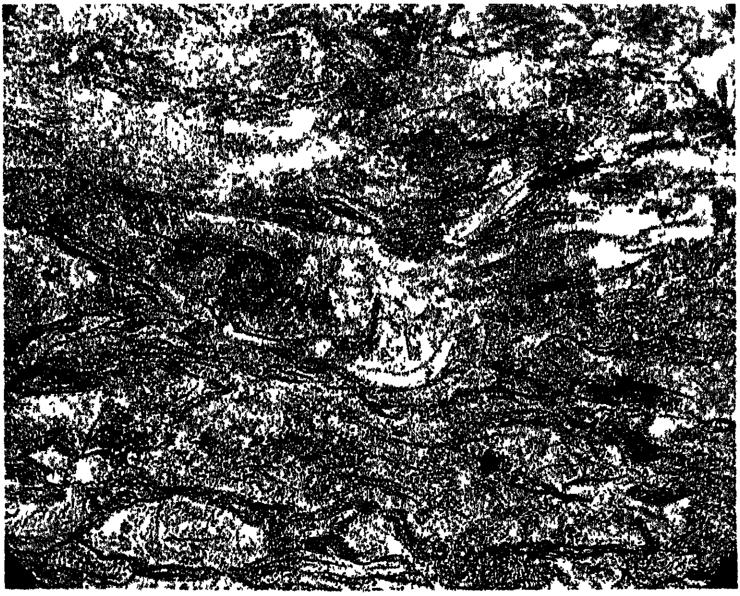

Fig 2 Flow structure in welded keratophyre tuff Specimen of matrix. Ordinary light (X 60).

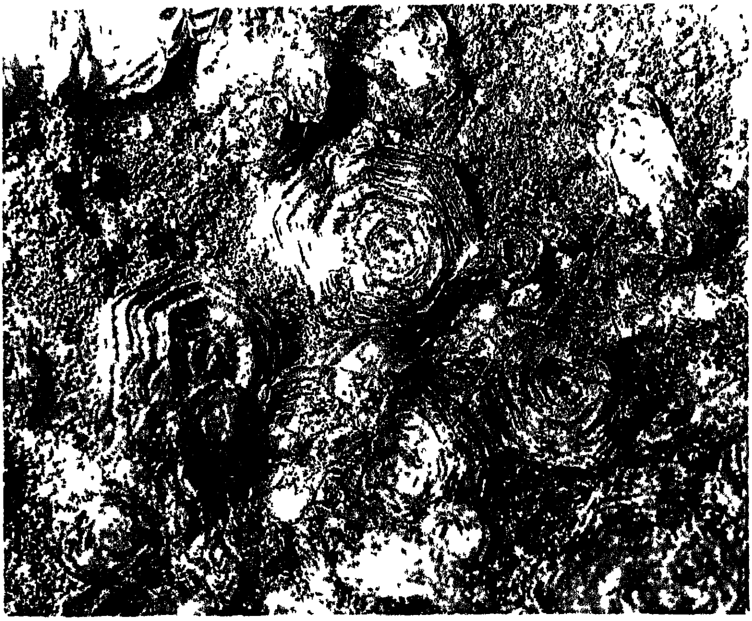

Fig 1. Zoned quartz crystals developed in vug in welded tuff spheroid Maximum diameter of individual crystals is about ½ inch

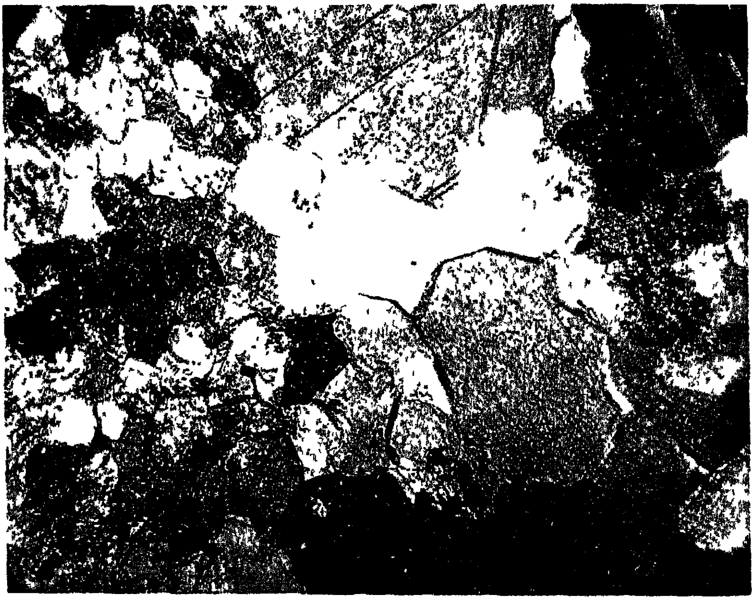

Fig 2 Tesselated calcite occurring in typical core of gray tuff spheroid Crossed Nicols (X 60)

by Ross (1942, p. 192) in connection with Oregon "thunder eggs." In other features of origin and internal structure, there appears to be little similarity between the two.

In none of the dozens of red spheroids cracked open was any megascopically visible core discovered. Nor is there any indication of radial structure in hand specimens. Weak color banding is displayed in a small percentage of the nodules.

Local concentrations of nodules having prominently developed concentric shell structure appear in limited areas along the outcrop. These nodules show a solid central mass, surrounded by a partial outer layer from 1/4 to 1/2 inch thick, separated from the central mass by an open space usually in the order of 1/4 inch thick. One-fourth to one-half of the core of the spheroid may be covered by this partial shell. Silica is commonly present in the vugs so formed. It occurs as mammillary deposits and as euhedral crystals of quartz projecting from both walls of the void, or, in some specimens, as crystals showing as many as ten concentric zones of growth (pl. 2, fig. 1). Some crystals are oriented with the *C* axis parallel to the wall of the cavity, but more crystals have the *C* axis normal to the wall. In the latter case, the zoned crystals show asymmetric growth, the greater development in about 75 percent of the specimens noted being toward a common locus on the circumference of the hemisphere which is the cavity wall. Each period of quartz deposition in the zoned crystals is marked by a distinct ridge separated from its neighboring ridge (in the same crystal) by a valley. The structure suggests alternate waves of deposition of quartz and some other mineral (calcite?) and subsequent removal of the other mineral, leaving the ridge-and-valley structure. The larger zoned crystals are as much as one-half inch in diameter and contain as many as ten distinct zonal ridges.

Other vugs contain deposits of powdery, black manganese dioxide, or of a very fine, white, powdery mineral which may be a zeolite. Rarely, all three minerals occur in a single vug.

In a ten-foot zone stratigraphically just below the spheroidal keratophyre tuff, almond-shaped vugs are abundant. These cavities range from one-half inch to six inches in length, and their height is approximately one-third the length. Some are empty; others are lined with any or all of the same minerals—quartz, manganese dioxide, or zeolite (?) — mentioned above.

Arrangement of filled or empty cavities appears entirely haphazard. It is not clear whether this zone is confined to a flow distinct from that in which spheroidal structures are well developed, or whether it is in the lower portion of the same flow. The almond-shaped vugs, however, are present, although in smaller number, in the zone of spheroidal development. Some serve as centers around which spheroids develop; but these spheroids tend to be weakly formed and rarely weather out of the matrix. This is in marked contrast to spheroids formed without determinable centers which weather out without loss of form.

A poorly defined joint system cuts across the keratophyre tuff, affecting both spheroids and matrix. No displacement was observed along these parallel breaks.

The area in which spheroids have been found in the gray tuffs is much smaller than that within the red. Spheroids in the gray tuff have been found only in one locality, a valley underlying a bluff formed by the keratophyre tuff. The area measures about 600 x 200 yards.

In hand specimen the tuff is light gray, fine-grained, and gritty. Prominent in the gray matrix are abundant xenoliths of red and purple volcanics, gray quartzite, and red shale. Shiny irregular flakes of a golden mica, probably altered biotite, are plentiful. Bleached zones adjacent to joints are prominent. Three sets of joints at right angles divide the rock mass into blocks, the edges of which are rounded by weathering. The effect, viewed from a little distance, is that of a pillow lava. Closer inspection reveals joint control.

Spheroids in the gray tuff are characterized by their relatively small size, one-fourth to three inches in diameter, by their hardness which exceeds that of the matrix, and by their irregular boundaries. The spheroids occur loose on the surface of the gray tuff and accumulate on small talus slopes below cliffs carved by gullying of the tuff. Most of them are not visible on the outcrop except where rimmed either by a zone of bleaching or by a zone of weak red stain. Comparable circular bleached areas, some of them rimmed by red stain, occur sparsely on the outcrop with no apparent associated spheroidal development.

Eighty percent of the spheroids examined have a calcareous core. The calcite is colorless, white, or rarely, gray, red, or black. In the remaining 20 percent of the spheroids, the cores consist of rounded fragments of quartz, quartzite, or sandstone. In none of the gray spheroids is a core completely missing.

The mode of occurrence suggests that the spheroidal masses are controlled by concentration of a cementing substance about a xenolithic nucleus. The assumption may be made that where tessellated calcite forms a core it represents a recrystallized fragment of limestone deposited contemporaneously with the tuff.

An objection to this interpretation is the abundance of xenoliths of shale, volcanic fragments, and limestone throughout the tuff, unaccompanied by the development of spheroidal forms. Another objection is local concentration of calcite grains in both tuffaceous matrix and spheroids.

Geological literature, insofar as can be determined, contains few pertinent descriptions of spheroidal masses in volcanic rocks. Structures that appear most nearly to resemble the nodules of either the red or the gray tuff phases are those in the figure accompanying Johannsen's (1937, pp. 300-301) discussion of variolites. The most detailed description of variolites is found in Rosenbusch (1908, pp. 1280-1288). These authors, however, confine the use of the term "variolite" to basic rocks in the basalt-diabase range on which spheroidal masses appear during weathering. The highly siliceous constitution of the two rocks under discussion makes the use of "variolite" seem inappropriate.

Stearns and MacDonald (1946, pp. 21 and 105) and Jaggar (1947, pl. 42-B, facing p. 467) have presented brief discussions of "ball lavas" formed either by accretion of basaltic lava around blocky cores on the surface of moving aa flows, or by formation of shells of glassy lava about rock kernels, again in basic lava. While Jaggar's photograph of the ball lavas of the Kau desert bears a superficial resemblance to the spheroids in the red keratophyre tuff of northern Canelo Hills, the absence of cores from the red tuff nodules, and the difference in composition of the lavas involved, make a close analogy appear unjustified.

Iddings (1887, p. 36) in his discussion of lithophysae in Obsidian Cliff, Yellowstone National Park, argues that incipient crystallization has taken place while the enclosing mass was still viscid. Anhydrous microlites of feldspar radiate from a center into a sphere of glass paste. Gases, principally water vapor, held as tiny bubbles within the magma aggregate into larger bubbles and act as superheated steam, facilitating the crystallization of all constituents of the magma with which they come in contact. After giving up all the contained gases, but before final crystallization of the glass paste, shrinkage occurs through reduction in volume, and the cracks commonly seen in lithophysae and in spherulites are formed. Iddings considers as evidence of pre-final solidification of the lithophysae the presence of tridymite, quartz, and fayalite on walls of cracks in the lithophysae. Iddings' discussion may bear on the formation of spheroidal structures in the keratophyre tuff of this paper, especially those showing concentric shells.

A closer analogy appears in Teall's (1888, pp. 336-338) discussion of "nodular felsites," in regard to which Teall says, citing earlier work by Cole and Bonney:

> "The 'nodules' lie in a felsitic matrix in which fluxion structure is often well defined. They may be solid or hollow. In the latter case, the cavities are often lined with quartz crystals. Sometimes the cavities have been entirely filled with quartz . . .
>
> "The nodules vary in size from that of peas to globes measuring four or five inches in diameter. They have sometimes been flattened by crushing. A soft black substance is occasionally found associated with them. It is sometimes arranged in concentric zones . . . "

Citing Iddings, Teall then summarizes Bonney's conclusions that the nodules have been formed in one of two ways: (1) by simple contraction and roughly concentric cracking during cooling, and: (2) by similar contraction which is determined by the presence of a cavity. In this case, as cooling proceeds, pressure of the gases within the cavity against the walls of the cavity rapidly diminishes. This reduction in pressure upsets a previous tendency toward equilibrium in the cooling glass paste, and contractile stresses result in formation of concentric cracks about the cavity.

Although Teal mentions association of "felsitic agglomerate" with the nodular felsites, it does not appear that the agglomerate contains any nodular structures. Nor have the present authors found any descriptions of comparable spheroidal structures in fragmental volcanic rocks.

OTHER OCCURRENCES

Other occurrences of spheroidal structures in volcanic rocks of Arizona will be cited, although they have not been examined in detail. Near Castle Hot Springs, about twenty miles east of Wickenburg, Brady[1] has found a basalt containing almost perfectly spherical masses which range in diameter to a maximum of about one inch. These spheroids are darker in color than the matrix and stand out distinctly on freshly broken surfaces of the rock. Examination under the microscope indicates that alteration controls these forms. Clay alteration is present in the spheroidal structures and not in the matrix.

Spheroids are known to occur in considerable numbers in the dacite near Superior mining area, Arizona, but apparently have not been discussed in the literature.

Well-rounded, white tuff balls have been sent to the Arizona Bureau of Mines from an unspecified locality near Wickenburg, Arizona. Their field occurrence is not described. Each has at its center an elongate core. The spheres range from about one-half inch to two inches in diameter.

An outcrop of gray volcanic agglomerate abounding in xenoliths of various rocks, as much as several feet in maximum section, occurs beside the Tucson-Ajo road ten miles southwest of Tucson. Small-scale quarrying operations have exposed the rock. Scattered through it appear spheroidal masses, centered about xenolithic or calcareous cores. Bleached zones, nearly circular on weathered surfaces, but not showing spheroidal development, are associated. The entire occurrence strongly suggests analogy with the gray tuff spheroids described earlier from northern Canelo Hills. The development of spheroidal structures, however, appears less advanced in this quarry than in Canelo Hills.

PETROGRAPHY

Under the microscope the matrix enclosing the red spheroids is seen to consist of a very fine-grained mass of quartz and feldspar, formed through the devitrification of glass. The absence of quartz phenocrysts and the presence of phenocrysts of secondary albite suggest the term keratophyre or keratophyre tuff. The characteristis texture of a welded tuff is strikingly shown by the matrix (pl. 1, fig. 2).

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[1] Brady, L F, Letter and specimen (1947).

Plagioclase phenocrysts in the matrix have been altered to albite. Spherulites are scattered throughout the matrix as small bodies, some well-formed and others quite irregular in shape. Typical shadow-crosses are seen under crossed nicols. Small veins of secondary quartz traverse the matrix in a few places.

Although the flow structure cuts across the line of demarcation at the edges of the spheroids, the texture is distinctly finer in the spheroids. "Ghosts" of the welded tuff texture are present in a few places. Aggregates of feldspar and quartz are aligned in the direction of flow movement, and are much more widely spaced than in the matrix where, apparently, devitrification has not been as complete.

Alteration is similar in spheroids and matrix. The few plagioclase phenocrysts have been altered to albite and veinlets of secondary quartz are present. Spherulites seem somewhat more abundant in the spheroids than in the matrix, and tend to clump together similarly, merging into one another.

Equatorial thin sections up to 2 x 3 inches were prepared to show internal relationships of the ball structures, but attempts to include in one section the boundaries of matrix and spheroids proved futile. As seen in the field, flow structure appears to cut directly across both matrix and spheroids. This has not been confirmed microscopically.

The only suggestion that a radial structure may be the controlling influence on the prominently weathering forms is a spacing of quartz-feldspar aggregates defining the primary flow texture (fig. 1). A suggestion of "paths" progresses radially from the centers of some of the sections examined toward the periphery. Under crossed nicols these darker "paths" are outlined by more intense light transmitted by the coarser quartz-feldspar aggregates between. Viewed with the polarizer alone, slightly greater relief of the aggregates causes them to stand out as oriented islands. The origin of this phenomenon is unknown.

POSSIBLE MECHANICS OF ORIGIN

On the assumption that the matrix surrounding the red spheroids is a welded keratophyre tuff formed by accretion of viscid lava particles, one possible explanation of these spheroidal structures is that they were extruded as larger and perhaps more fluid masses than the rest of the tuff. By virtue of their greater fluidity they might have been more completely bonded together at the time of ejection. Immediate burial by another layer of lava or by additional fluid ash material and subsequent movement of the whole mass down a slope would produce flow structure in matrix and spheroids alike, with textures differing according to the characteristics of the materials involved. One would not, perhaps, expect to find in structures formed in this manner the high degree of sphericity exhibited by the red spheroids, as partial rolling would tend to flatten the larger bodies.

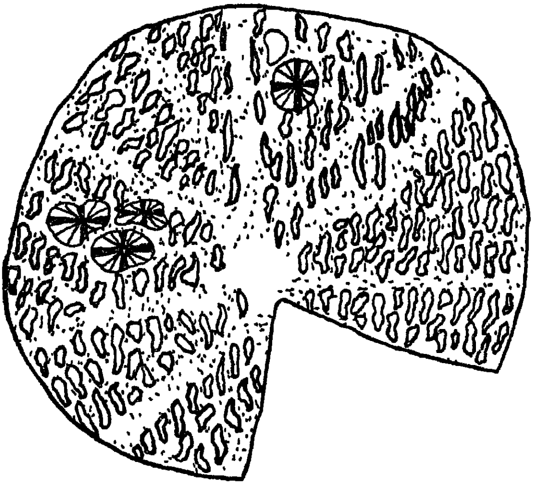

Figure 1 Diagrammatic representation of "paths" developed among oriented quartz and feldspar grains in groundmass of welded tuff spheroids, suggesting radial development.

Re-heating of a welded tuff or flow might tend to cause more highly siliceous areas to become less viscous sooner than the more basic remainder. Re-solution of the tuff starting from more highly siliceous centers would tend to change the character of the rock radially away from these loci. The assumption must here be made that the welded tuff, as laid down, was not completely homogeneous; areas more siliceous than the main mass were scattered throughout the flow, and after the tuff had become compacted, welded, and perhaps cooled (but not crystalline), another thick flow moving over it would supply the heat necessary to start the melting only in the more siliceous areas.

Such incipient re-solution could probably destroy the flow structure of the welded tuff to the extent that it took on a finer-grained appearance, and "ghosts" of the coarse welded tuff texture might be left here and there.

If the silica present were in the form of quartz, this process probably would not take place, as the amount of heat required to melt the quartz crystals would be greater than that needed to melt the glassy matrix.

The presence of small amounts of water locally in the tuff would tend to lower the melting point greatly and have the same effect as more highly siliceous areas in remelting the tuff.

GRAY TUFF

The gray spheroids are seen under the microscope to be a typical tuff consisting of a fine-grained matrix with sporadic glass shards. Devitrification has been nearly complete. Most of the sections cut have masses of interlocked calcite grains completely occupying the central portion (pl. 2, fig. 2). Some of these calcite masses show evidence of zoning as if the calcite had been introduced from the outside to fill a vug or to replace some other material. Others show no evidence of this.

Both the matrix tuff and the spheroidal masses contain a great deal of calcite scattered at random throughout. Some of the spheroids have masses of calcite as large as their cores out near their peripheries.

The presence of "valleys" between layers of silica in the zoned quartz crystals mentioned earlier may be explained by calcite deposition alternation with the waves of silica deposition, provided that the calcite layers were subsequently removed by solution. The close proximity in the field between the limited area in which zoned crystals have been found and that in which the gray tuff nodules occur suggests a related secondary origin for both. No additional lines of evidence supporting such a hypothesis have been found. The writers feel that for the keratophyre nodules, however, the conclusion that they are of primary origin better satisfies the known facts.

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THE ULTRABASIC ROCKS OF MIDDLE RIVER RANGE, B.C.*†

H. W. LITTLE

ABSTRACT The ultrabasic rocks of the Middle River Range of British Columbia consist of sills and stocks of dunites and peridotites, with minor pyroxenite and norite. Field evidence indicates that the ultrabasic rocks were probably derived from a magma of peridotitic composition. It is believed that the solutions which brought about partial to complete alteration of the ultrabasic rocks to serpentine were derived in part from the ultrabasic magma and in part from the intruded sedimentary rocks. The migration of solutions from the country rock into the intrusives results in cooling of both the country rock and the intrusives and is responsible for the low grade of metamorphism surrounding these ultrabasic bodies.

INTRODUCTION

As a result of the need for chromite and other war minerals, field parties were organized by the Geological Survey of Canada in 1942 to prospect for this mineral. Dr. J. E. Armstrong was assigned to the ultrabasic massif of the Middle River Range, and the writer was in charge of a sub-party in the west half of the area. This provided an unusual opportunity to study the ultrabasic rocks in great detail, and although it was not intended that an official report be published, such a wealth of additional knowledge of this area was obtained that the writer feels a report is justified.

LOCATION OF THE AREA

The Middle River Range lies between Trembleur Lake and Takla Lake in central British Columbia. The area mapped extends from latitude 54° 48′ to 55° 01′ and from longitude 125° 08′ to 125° 30′. It is approximately 300 square miles in area. (fig. 1)

PREVIOUS WORK

The first reference to any part of the region mapped was by Camsell (1916, pp. 70-75) who made a rapid reconnaissance trip along part of Trembleur Lake and Middle River.

In 1937, Armstrong (1941), while carrying out the reconnaissance mapping of the West Half of the Fort Fraser area, discovered and outlined the ultrabasic rocks. In 1940, he mapped the rocks of the adjoining Takla sheet (Armstrong, 1946), north of latitude 55° 00′.

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* Part of a thesis submitted to the University of Toronto in partial fulfillment of the requirements for the degree of Doctor of Philosophy

† Published by permission of the Chief Geologist, Geological Survey of Canada

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The writer takes great pleasure in thanking all those who have given their assistance in the preparation of this work. He is especially grateful to Dr. J. E. Armstrong for the use his field notes, specimens, and maps, and for the loan of many thin sections. In the west half of the area Mr. D. Carlisle rendered very efficient assistance to the writer during the field season.

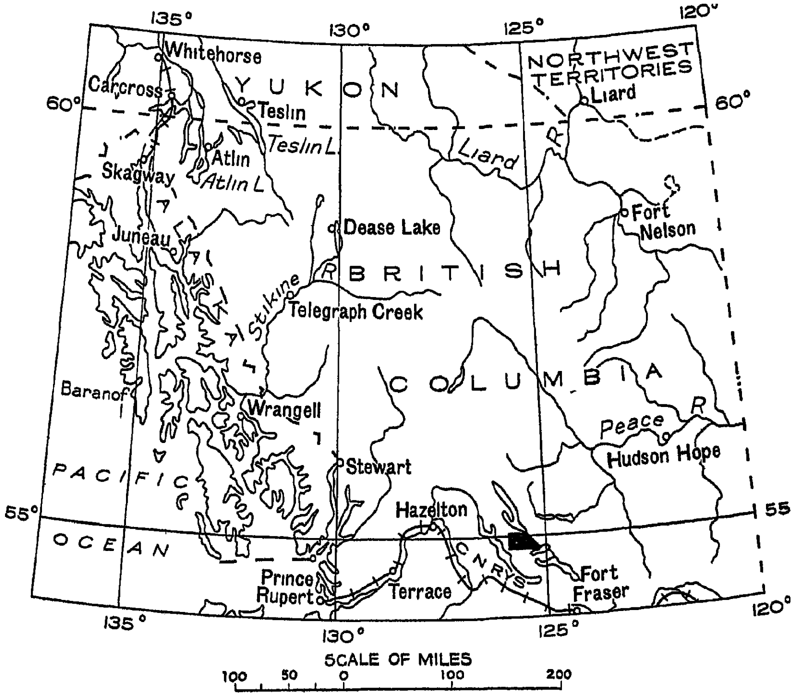

Fig 1 Location of the area.

At the University of Toronto, Drs. E. W. Nuffield, F. G. Smith and W. W. Moorhouse offered many valuable suggestions and other material contributions.

Although the views of the writer differ from those of Drs. N. L. Bowen and T. F. W Barth, he is greatly endebted to them for a critical review of the manuscript.

GENERAL GEOLOGY

The oldest rocks in the area are presumably the Cache Creek series of Permian age, representing a comformable succession, not less than 22,000 feet of interbedded sedimentary and volcanic rocks and their derived metamorphic types. There are three divisions; a lower group of massive recrystallized limestone, a middle group of slate and argillite, with minor limestone and greenstone and related schists, and an upper group of greenstone, with lesser argillite, slate and minor limestone. All members strike northwest and have been tightly folded, the dips normally being steep. Regional metamorphism of the series is widespread and quite intense.

The Permian rocks have been intruded by Mesozoic (?) stocks, sills, and dykes, ranging in composition from ultrabasic to acidic. The ultrabasic bodies, which are older, are in general highly serpentinized and sheared, and have, in places, been altered to carbonate-talc-quartz rocks. These rocks were originally mainly peridotites and dunites, with minor pyroxenite and norite. The acidic bodies are composed of soda granite, quartz porphyry, and minor related acidic and basic dyke rocks.

The extreme western part of the area is underlain by a great thickness of sedimentary and volcanic rocks of the Takla group, which is Jurassic and possibly Upper Triassic in age. These rocks are intruded by acidic stocks and dykes, but in no place are they cut by rocks of ultrabasic composition.

A widespread mantle of Pleistocene and Recent material which covers much of the area was not mapped.

ULTRABASIC AND RELATED INTRUSIVES

The oldest group of intrusive rocks in the area consists of peridotites, dunites, pyroxenites, and norites in descending order of abundance. Owing to the small size and complexity of the segregated bodies, most of them cannot be separated on the map. For this reason areas have been classified on the map according to the rock type which predominates in that particular locality.

In general, nearly all of these interrelated rock types are considerably altered—the olivine to serpentine and magnetite, the pyroxenes to amphiboles, serpentine, and magnetite, and the plagioclase, which appears only in the norite, to saussurite.

The ultrabasic complex is estimated to consist of more than 75 per cent peridotite in the east half of the area, but in the west half where differentiation into dunite and peridotite is more marked, it consists of only about 65 per cent peridotite. Those peridotites which have not suffered hydrothermal alteration are dark green, medium-grained rocks, usually containing from 10 to 60 per cent pyroxene, the average being nearer the lower figure. In addition to pyroxene, the peridotites contain olivine, with minor amounts of picotite, chromite, or magnetite.

The peridotites are predominantly of the saxonite or harzburgite type. A few of the wehrlite type were observed, but none of the lherzolite type. The pyroxene crystals are generally much larger than those of olivine, which are sometimes encased

Table of Formations

Age	*Name*	*Description*
Pleistocene and Recent		Recent alluvium and glacial drift (moraines and erratics).
Post-Jurassic and (?) Pre-Upper Cretaceous		White soda granite, pink biotite granite and related dyke rocks.
	Tsitsutl stock	Quartz porphyry and related dyke rocks.
Jurassic and (?) Upper Triassic	Takla group	Andesite and basalt flows, tuffs and breccias, with interbedded sandstone, shale and conglomerate
Post-Permian and Pre-Upper Triassic	Trembleur intrusions*	Peridotite and dunite, with minor pyroxenite and norite, and related alteration products — serpentine and carbonate-quartz-talc rocks.
Permian	Cache Creek series	Andesite and rhyolite flows and tuffs, with lesser argillite and slate and minor limestone.
		Argillite and slate, with minor limestone and greenstone and related schists.
		Massive dark grey recrystallized limestone.

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* The name "Trembleur intrusions" has been applied by J E. Armstrong to many bodies of ultrabasic rocks of probable pre-Upper Triassic age that are exposed in the Fort St James Map-Area, B C., Geol. Surv, Canada (memoir in press).

in pyroxene, giving rise to poikilitic texture. In this case the pyroxene is often altered to serpentine, whereas the smaller olivine crystals are virtually unaltered. Where the olivine crystals are larger, and the proportion of pyroxene is less than average, the olivine appears to be as readily altered as the pyroxene. Olivine is usually altered to antigorite along regular fractures in the crystal. Orthorhombic pyroxene on the other hand frequently appears to have been partly altered to thin laths of tremolite along the prominent cleavages, and later the remainder was altered to antigorite. Augite, however, was altered partially to uralite in only one case, and never to serpentine.

The dunites as a rule are darker in colour than the peridotites and somewhat finer in texture. On the weathered surface they present a buff-coloured, granular appearance. The dunites appear in sills, and as irregular zones or bands between similar zones of peridotite in thick, differentiated sills, or as irregular masses in the batholithic body underlying Mount Sidney Williams.

The dunites may be seen in thin section to consist of coarse grains of subhedral olivine, usually considerably altered to antigorite, and minute grains of picotite, chromite, or magnetite are common.

The pyroxenites are in general rather coarser in grain than the peridotites and present a fibrous appearance. They are green in colour and may be almost fresh, or they may be nearly completely altered to serpentine with or without the presence of uralite or tremolite.

Thin sections show that the pyroxenites are composed essentially of large crystals of pyroxene more or less altered to antigorite and magnetite. Partial alteration of the pyroxene to hornblende or tremolite may also have taken place.

Pyroxenites occur as lenticular masses of limited size either in peridotite or dunite, or in the pyroxenite-norite complexes which occur in the east half of the area. They form also extremely coarse-grained branching dykes, varying in width from a fraction of an inch to two feet, cutting across the primary banding of the peridotites.

The norites are highly altered dark green medium-grained rocks. The presence of plagioclase feldspar distinguishes them from other rocks of the ultrabasic complex.

The rock in thin section is observed to consist of pyroxene, usually completely altered to amphibole, magnetite and epidote, and of plagioclase which is highly saussuritized. In a few in which the pyroxene is sufficiently fresh to be identified, it was found to be hypersthene Primary magnetite is present in some specimens. The norites occur only in the east half of the area either in the pyroxenite-norite complexes, or as altered dykes up to 30 feet in width cutting the peridotites on Mount Sidney Williams.

Internal Structures of the Ultrabasic Rocks. The relationships between the different rock types in the ultrabasic complex are of paramount importance in an analysis of their origin. Within the northern part of the ultrabasic batholith, that part underlying Mount Sidney Williams, dunite forms irregular masses (fig. 2) varying from a few square feet to several thousand square feet in area within the peridotite. These masses have no regular distribution, and are as numerous near the borders as they are in the interior of the batholith. The peridotite-dunite contacts are gradational across 2 to 4 inches. It has been estimated that within this moderately altered part of the batholith dunite probably exceeds one quarter of the total mass.

From the peak of Mount Carlisle eastward to their contact with greenstone, the ultrabasic rocks consist of thick differentiated alternate layers of dunite and peridotite, the latter predominating. Within this locality there are six sill-like differentiated zones of dunite in the peridotite. These zones, which strike about north-south and dip vertically, are from 50 feet to 100 feet wide and can be traced longitudinally for several hundred feet, their extremities either terminating in precipitous cliff faces or else disappearing under a cover of drift. Within this massive dunite, the olivine is often coarse-grained and considerably serpentinized. The dunite has a typical yellow-brown sandy weathered surface. Chromite is common as an accessory, varying from 1 per cent to more than 5 per cent. In the latter case it forms nodular masses up to three-eighths inch in diameter, from which thread-like veinlets of chromite extend for several inches. It was noted in the field that in these zones the chromite content increases appreciably at lower elevation; this gradation is therefore perpendicular to the walls of the zones.

In the stock three miles south of Pyramid Peak and in the broad sill south of Tsitsutl Mountain, primary banding of the peridotite is striking, and is clearly visible from considerable distances. Detailed examination of the layers reveals that they are usually about 2 inches thick, consisting of peridotite containing about 10 per cent pyroxene, separated by peridotite

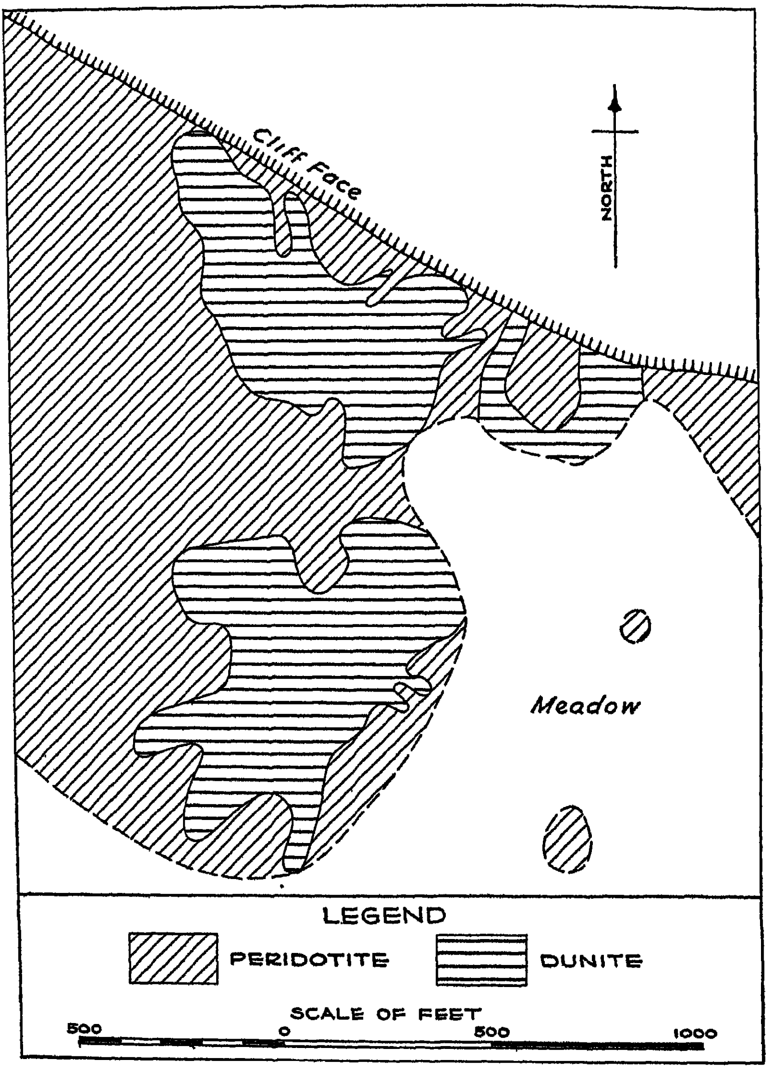


Fig. 2. Irregular masses of dunite in peridotite on Mt Sidney Williams.

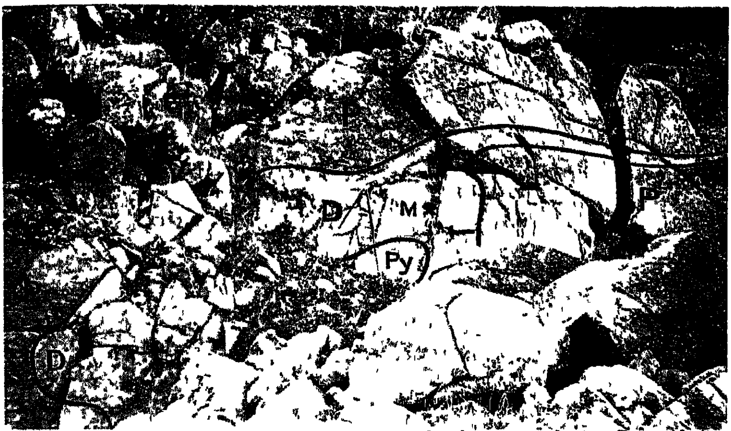


Fig 1 Irregular dunite mass cutting across primary banding in peridotite. P - peridotite, D - dunite, Py - pyroxenite, M - magnetite

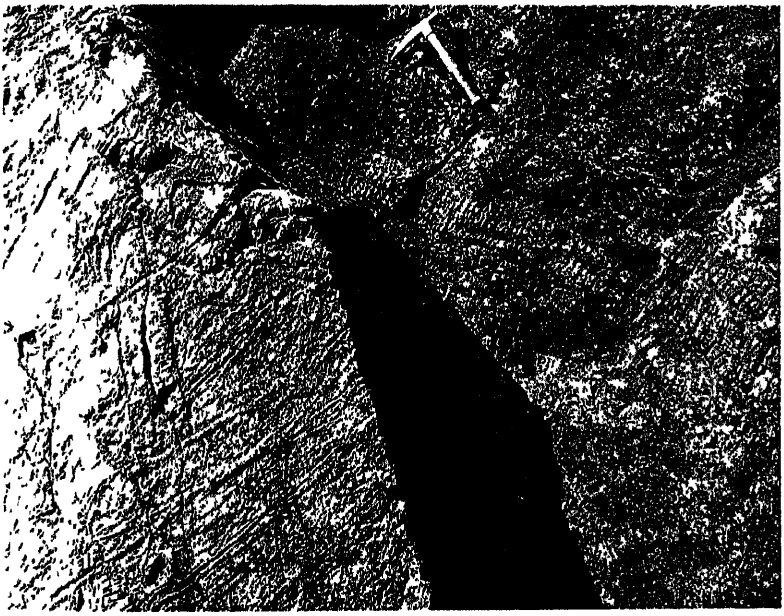


Fig. 2 Pyroxenite dykes cutting peridotite on Mount Sidney Williams

with about 40 per cent pyroxene, and approximately 1 foot in thickness. The base of each narrow band is sharp, but the top is gradational across ½ inch, which is indicative of crystal settling. Frequently the pseudostratification of the peridotites is gently curved, suggesting a bending while still plastic, and tiny fractures show displacements of several inches, which gradually decrease to zero a few feet distant on the outcrop faces. The outcrops of banded peridotite appear to form separate blocks whose attitudes vary from one to another as though separated by minor faults.

In the thick sill dunite, in addition to occurring as segregated layers, forms small, irregular bodies with long apophyses into the peridotite. The contacts, however, are seldom sharp, and never knife-edge, possibly owing to reaction between the two rock types. In dunite of this type, magnetite is a common accessory, rather than chromite. One such body is noteworthy for its peculiar structure. It consists of a bulbous mass of fine-grained dunite about 5 feet in diameter, which grades toward the centre into a coarse pyroxenite. A small mass of magnetite lies just above the pyroxenite. An apophysis of dunite nearly 15 feet long and a few inches wide extends from the bulbous mass across the primary banding of the peridotite host-rock (pl. 1, fig. 1).

On Mount Carlisle, in addition to occurring as broad zones, dunite forms lenticular masses up to 5 feet or more wide and at least 10 feet long. The non-parallel walls are in rather sharp contact with the peridotite so that they appear to be later in age than the latter.

Dunite also occurs as sills such as the one which lies about 1 mile west of West Peak. While no peridotite occurs in these sills, lenticular masses of pyroxenite up to more than 100 feet in length are often present. They are rather coarse in texture, the majority of pyroxene crystals exceeding ¼ inch in length. The pyroxenite grades within a few inches through a peridotitic phase into dunite.

In the broad sill south of Tsitsutl Mountain, where the pyroxene content of peridotite is high (60 to 70 per cent), irregular xenoliths of peridotite containing perhaps 30 per cent pyroxene are often present. These xenoliths stand out clearly on the outcrops because of their fine-grained texture and particularly because of the rusty weathering of the surface in contrast to the typical orange-weathering of the peridotite that encloses them.

Near Mount Sidney Williams several lenses of pyroxene-rich peridotite up to 500 square feet in area were observed in the peridotite. These have gradational contacts across a few inches with normal peridotites.

Pyroxene-rich peridotite and pyroxenite occur also in dykes which intersect the ultrabasic complex in the same locality. These dykes are irregular and branching, varying in width from a fraction of an inch to more than a foot (pl. 1, fig. 2). Similar dykes as much as two feet wide were observed near Mount Carlisle. They consist of extremely large crystals of pyroxene which grow perpendicular to the walls of the dykes. In such dykes chromite is present in minor amount and is always clearly interstitial. In one dyke pyroxene was brecciated and chromite forms a sparse cementing material.

One of the bodies underlying West Peak is composed almost entirely of a norite-pyroxenite complex. A similar complex occurs near Trembleur Lake and is completely surrounded by carbonate-quartz-talc rock. The norite appears to be relatively fresh in the hand specimen, and consists of about one-half white feldspar, the remainder being short, prismatic crystals of pyroxene. It is intimately associated with small masses of very coarse-grained pyroxenite. Both of these rocks are very little affected by the carbonate solutions which have replaced the peridotites and dunites in the vicinity. Thin sections of the rocks reveal, however, that the plagioclase has been somewhat saussuritized although the hypersthene is little altered. Several small outcrops of pyroxenite occur elsewhere within the carbonate-quartz-talc rocks.

In the vicinity of Mount Carlisle north-south fault zones 25 to 100 feet wide traverse the serpentinized ultrabasics, displacing acidic and basic dykes that cut the serpentine. They are recognizable as zones of schistose serpentine in which slickensides are common. In many cases ellipsoids of massive serpentine up to several feet in diameter occur in the zones. The schistose serpentine throughout the faults is usually light green and sometimes this effect is noticeable on the outer half-inch of the ellipsoids. Such light-green serpentine does not weather on the surface to the typical orange or yellow-brown of the normal dark serpentine.

Serpentinization of the Peridotites, Dunites, and Pyroxenites. The peridotites, dunites, and pyroxenites have undergone partial to complete alteration to serpentine, so that in thin section it is often impossible to determine the original constituents of the rock. In hand specimens, however, pseudomorphs of serpentine after orthorhombic or monoclinic pyroxene are easily recognized by cleavage faces on the broken surfaces, or by outlines of the crystals on the weathered surfaces. The serpentine is mainly antigorite and is generally olive green, except in fault zones where it is often light green, or in contact with acidic dykes where it assumes a characteristic chocolate brown colour. The fibrous variety of serpentine, chrysotile, is often observed in minute fractures and in fissures, where it may attain a length of 10 inches or more. No chrysotile of commercial quality has been found in the Middle River Range; that which does occur is much too brittle.

The progress of serpentinization has been described in detail by Armstrong (1940, pp. 21-32), and the writer does not need to reiterate his statements Certain additional observations were made during a study of thin sections and these are as follows:

1. In general, dunite is most highly serpentinized, the olivine being generally almost completely altered to serpentine; the original rock is recognized largely by means of its characteristic texture. Serpentinization of peridotite varies from virtually no alteration to almost complete alteration. Where the texture of peridotite is equigranular, olivine is invariably more completely altered than pyroxene, but where the texture is poikilitic the large pyroxene crystals are often completely altered to serpentine, while the smaller olivine crystals which are encased in the pyroxene are practically unaffected. In pyroxenites the pyroxene is sometimes quite fresh, but more often it is partly altered to amphibole and serpentine.

2. Previous to serpentinization, orthorhombic pyroxenes have in many cases partially undergone what is apparently deuteric alteration to amphibole. This effect took place along the cleavages of the pyroxenes. Where serpentinization accompanied or followed this process, acicular crystals of tremolite with a common orientation mark the positions formerly occupied by pyroxene.

Facts pertaining to serpentinization of the ultrabasic rocks of the Middle River Range are enumerated below:

1. Serpentinization is invariably more complete in small ultrabasic bodies, especially those of dunite, than in those of larger size.

2. Serpentinization is more intense within fault zones than in unsheared ultrabasic rocks. While this fact at first sight appears to favour an hydrothermal source for serpentinizing solutions, it may be explained by the fact that those parts of the ultrabasic bodies that have been most highly serpentinized were the most incompetent and therefore tended to localize shears resulting from stress. This would be true particularly if serpentinization were accompanied by a decrease in volume.

It has been noted previously that in the majority of shear zones in the ultrabasic rocks, there is present a spotted light-green serpentine which does not exhibit the usual orange-weathering limonitic coating of normal dark green serpentine. Schistose serpentine is as a rule completely altered to the light-green variety, but where ellipsoids of massive serpentine are present in a shear zone this alteration does not penetrate more than one-half inch from the surface of the ellipsoids. Such bright green serpentine is believed by the writer to be the effect of leaching of the iron oxides, which have been carried away. This removal of iron oxides was apparently accomplished by hydrothermal solutions ascending the fault zone, since in one locality a network of tiny calcite veinlets has been introduced into fractures in the serpentine.

3. The distribution of serpentinization bears no relation to the proximity of younger acidic intrusives that traverse the ultrabasic bodies. It is therefore evident that serpentine was not produced by hydrothermal solutions emanating from the acidic intrusives themselves.

Armstrong (1939, pp. 77-78) has reviewed the development of the various theories that have been proposed to explain the processes of serpentinization of ultrabasic rocks. He has shown that the more recent investigators favour an autometamorphic process in which serpentinizing solutions accumulate within the ultrabasic reservoir and migrate to the consolidated portions, altering the olivine and pyroxene to serpentine. Hess (1933, pp. 634-657; 1938, pp. 321-324) has further modified this conception by suggesting that the serpentinizing reaction was not only auto-metamorphic, but deuteric in the most restricted sense. He also offers the suggestion that if water-bearing sediments have been intruded by ultrabasic rock the former will probably contribute additional water to the serpentinizing solutions. He apparently believes that this contribution is of minor significance, since he suggests that the ultrabasic magma contains from 5 to 15 per cent water, the latter quantity being more than sufficient to bring about complete serpentinization of dunites and peridotites.

Morey and Ingerson (1938, pp. 217-225) have shown that at high temperatures and pressures, water and alkali silicates are mutually soluble in all proportions. Goranson (1938, pp. 71-91) has found that the maximum solubility of water in albite melts and in glasses of Stone Mountain granite is approximately 10 per cent. Smith (personal communication) believes that the solubility of water in silicate melts bear a direct relation to the alkali content of the melt, so that by extrapolation of the solubility curve, the maximum solubility of water in a melt of ultrabasic composition would probably not exceed 2 or 3 per cent. If this were the case, the serpentinizing solution could not have originated entirely within the ultrabasic magma.

Field evidence indicates that the injection of the ultrabasic magma began when the sediments that they intruded were still flat-lying, or nearly so. It is therefore probable that the sediments contained a considerable amount of water, a large proportion of which may have contributed to the serpentinization of the ultrabasic rocks. This hypothesis most clearly explains why the smallest bodies of ultrabasic rocks are the most completely serpentinized.

Hess (1933, p. 651) and Cooke (1937, p. 67) have listed several of the most probable chemical reactions by which dunites, peridotites, and pyroxenites may be altered to serpentine. In such reactions, where serpentinizing solutions have been introduced there results an increase in volume of from 25 to 84 per cent. If the serpentinizing solutions were present in the original magma, only small volume increases, and in some cases, volume decreases occur during serpentinization. If a portion of the solutions was derived from the magma and the remainder from the country rock, it is approximately equivalent to assuming that all the solutions came from the magma, since the shrinkage of the country rock upon dehydration would tend to compensate for the expansion brought about by serpentinization. Therefore it is unlikely that serpentinization brought about a very great expansion of the rocks as a whole, and this expansion would express itself in faulting and other small movements. The serpentinizing solutions consisted mainly of water, derived in part from the ultrabasic magma itself, in part from the intruded rocks. In the latter case, alkaline silicates and carbonates were probably present in aqueous solution.

Steatitization of Dunites, Peridotites, and Pyroxenites. Steatitization of the serpentinized dunites and peridotites is so extensive in the east half of the Middle River Range that the altered rocks have been mapped as a separate petrographic unit. From a study of the map it can be seen that the most extensive alteration of this type occurs around the periphery of the ultrabasic batholith, particularly toward the southeast. The best exposures occur along Trembleur Lake and the ridge to the northwest which parallels the shore.

The hydrothermally altered rocks are greenish grey to buff and consist of buff-coloured carbonate, pale green talc, and white quartz in various proportions. Euhedral to subhedral grains of magnetite and chromite are often present as accessory minerals. Occasionally a chrome-bearing, apple-green mica (mariposite) is observed, and euhedral crystals of pyrite are less common. Armstrong (1940, p. 32) has reported chalcopyrite as an accessory mineral, but it was not seen in specimens examined by the writer.

The carbonate-talc-quartz rocks replace serpentine most easily; slightly serpentinized pyroxenites are least affected.

Armstrong (1940, pp. 30-31) has stated that the hydrothermally altered ultrabasic rocks can be divided into two distinct types: a carbonate-quartz-mariposite type and a carbonate-talc type. This implies that where quartz or mariposite are present, talc should be absent. The writer believes that no such classification is possible. Where alteration is slight, small quantities of talc or carbonate or both may be found in the rock. Where the alteration is more advanced, quartz may also be present. Where replacement of the original peridotites or dunites is complete or nearly so, the rock consists of at least three minerals: carbonate, talc, and quartz. The chrome-mica mariposite was found by the writer in only two thin sections, where it is associated with quartz and carbonate; talc is lacking although it is abundant in specimens taken nearby.

The carbonate which is abundantly present in most of these altered rocks is buff-coloured and weathers to a yellow-brown coating of limonite. Two samples of the carbonate were analyzed by the writer, who found that they are entirely lacking in calcium and manganese. The composition lies between that of magnesite, $MgCO_3$ and mesitite, $MgFe\ (CO_3)_2$.

This type of alteration was most probably brought about by the attack of dilute carbonate solutions upon serpentine. The chemical reactions may be expressed in this manner:

$$\underset{\text{serpentine}}{2H_4Mg_3Si_2O_9} + 3CO_2 = \underset{\text{talc}}{H_4Mg_3\ (SiO_3)_4} + MgCO_3 + \underset{\text{magnesite}}{3H_2O}$$

$$\underset{\text{serpentine}}{3H_4Mg_3Si_2O_9} + 3CO_2 = \underset{\text{magnesite}}{3MgCO_3} + \underset{\text{quartz}}{SiO_2} + 2H_2O$$

The iron, which is always present in the carbonate, must have been derived from the magnetite everywhere present in the serpentine. The water released by both reactions would serve to serpentinize the ultrabasic rocks above, which had not been completely altered to serpentine, and prepare them for alteration to carbonate, talc, and quartz.

Both of the above reactions have taken place during serpentinization, although locally one or the other may predominate. Temperature may be the chief factor governing these reactions, but a glance at the equations suggests that where carbon dioxide is abundant the formation of talc and carbonate will be favoured, rather than that of carbonate and quartz. From a study of the thermo-electric potential of the euhedral pyrite crystals present in two hand specimens of carbonate-quartz-talc rock, F. G. Smith determined their temperatures of formation to be approximately 450° and 500° C. respectively (1943, pp. 519-523).

The contacts of all ultrabasic rocks have undergone steatitization. This is probably due to the migration of hydrothermal solutions induced in the invaded rock by the intrusive rock. Such solutions would no doubt contain carbonates which would react with the chilled margins of the ultrabasic bodies, where the temperature may be favourable to such reactions. Similar steatitization penetrating several inches into ultrabasic rocks was seen where they have been invaded by acidic dykes, but this form of alteration has not always taken place.

An underlying magma is postulated as the source of the hydrothermal solutions that brought about the steatitization of the major part of the ultrabasic batholith. The composition of this magma is not known, since no known off-shoots from it are exposed in the locality.

Metamorphism of Invaded Rocks by Ultrabasic Bodies. One of the most controversial subjects with regard to the origin of ultrabasic rocks is the general lack of metamorphism surrounding intrusives of this composition. The first step in an analysis of this problem is to determine the temperature at which the ultrabasic rocks were intruded. Bowen and Schairer (1935, pp. 151-217) have discussed the system MgO-FeO-SiO_2, and have shown that magmas having the composition of the ultrabasic rocks of the Middle River Range must be almost completely crystallized at 1500° C. in a dry melt. The authors point out, however, that the presence of a small amount of Fe_2O_3 in the system complicates the phase diagram greatly. The effect of other constituents is unpredictable, but it seems probable that the addition of potassium, sodium, calcium, and aluminum to the system would lower the melting intervals of the members by at least 100° C., and the addition of 2 to 3 per cent water, with small amounts of boron, fluorine, and chlorine would have an even greater effect. The effect of pressure at a depth of 5 miles would be to increase the melting point of peridotite about 50°.

The writer postulates that the ultrabasic magma was injected at a temperature of about 1200° C. This figure agrees with the estimate derived from field evidence by Ingerson (1935, pp. 422-440) in the Trout River area, Newfoundland.

On the assumption that the temperature of the invaded rocks was 100° C., the maximum temperature reached by any point in the invaded rocks surrounding a sill 400 feet thick was calculated by this writer (Ingersoll and Zobel, 1913, p. 70). The results are plotted in figure 3. It is apparent that the temperature of the invaded rocks more than 15 feet from the

LEGEND

MESOZOIC

UPPER TRIASSIC (?) AND JURASSIC

TAKLA GROUP

⋰ Porphyritic and amygdaloidal lavas (andesite, rhyolite, basalt), argillite, sandstone and conglomerate

PALÆOZOIC

PERMIAN

CACHE CREEK GROUP

▤ Andesite (greenstone) and rhyolite, minor argillite and limestone

▨ Argillite and slate, minor andesite (greenstone) and limestone

▥ Massive limestone, minor argillite and slate

MESOZOIC

✚ Quartz porphyry

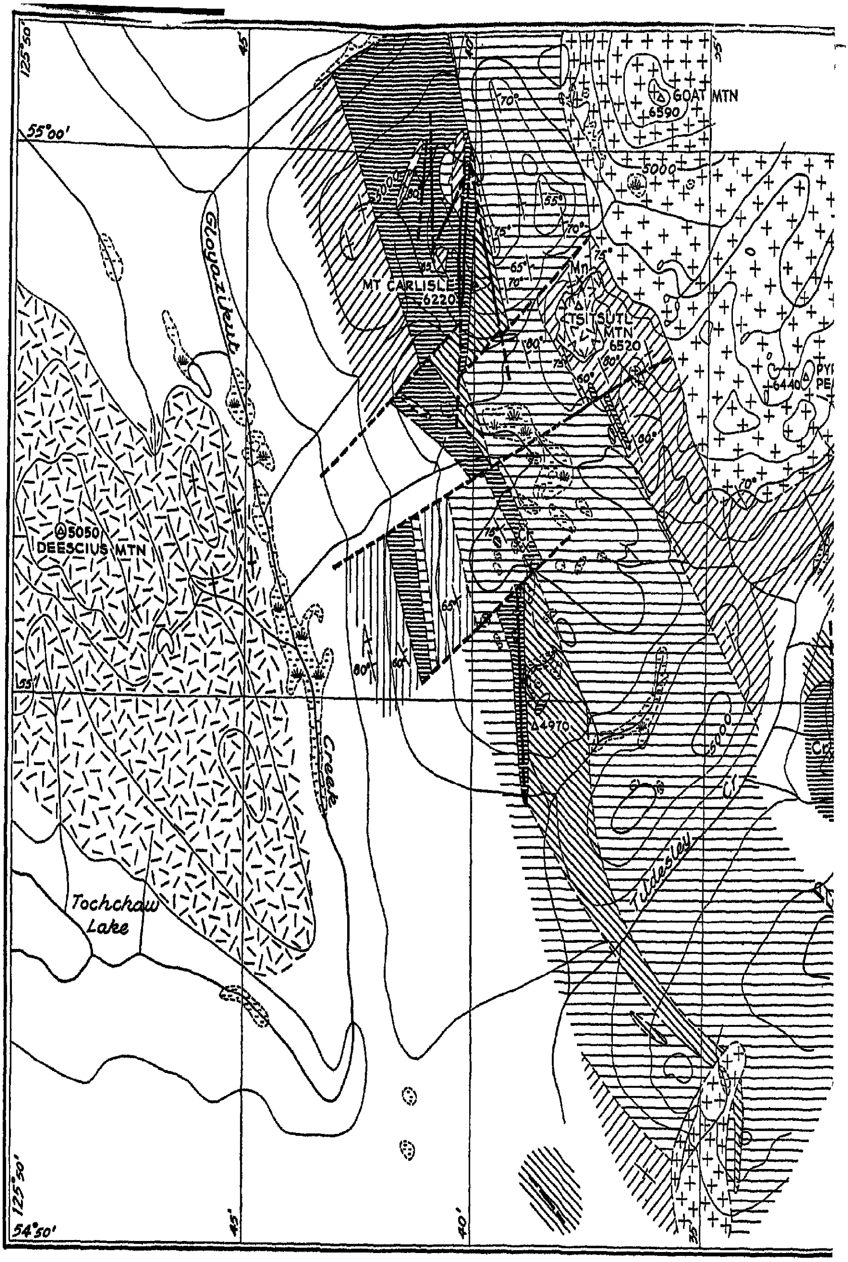
∨ Biotite granite

■ Norite-pyroxenite complex

▧ Predominantly peridotite, with dunite and minor pyroxenite

☰ Predominantly dunite, with peridotite and minor pyroxenite

☐ Carbonate-talc-quartz rocks (altered ultrabasic rocks)


125°50'
45'
40'
35'
55°00'
55'
GOAT MTN
6590
5000
MT CARLISLE
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6520
PY
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6440
4970
5000
DEESCIUS MTN
5050
Gloyazikut
Creek
Tildesley Cr
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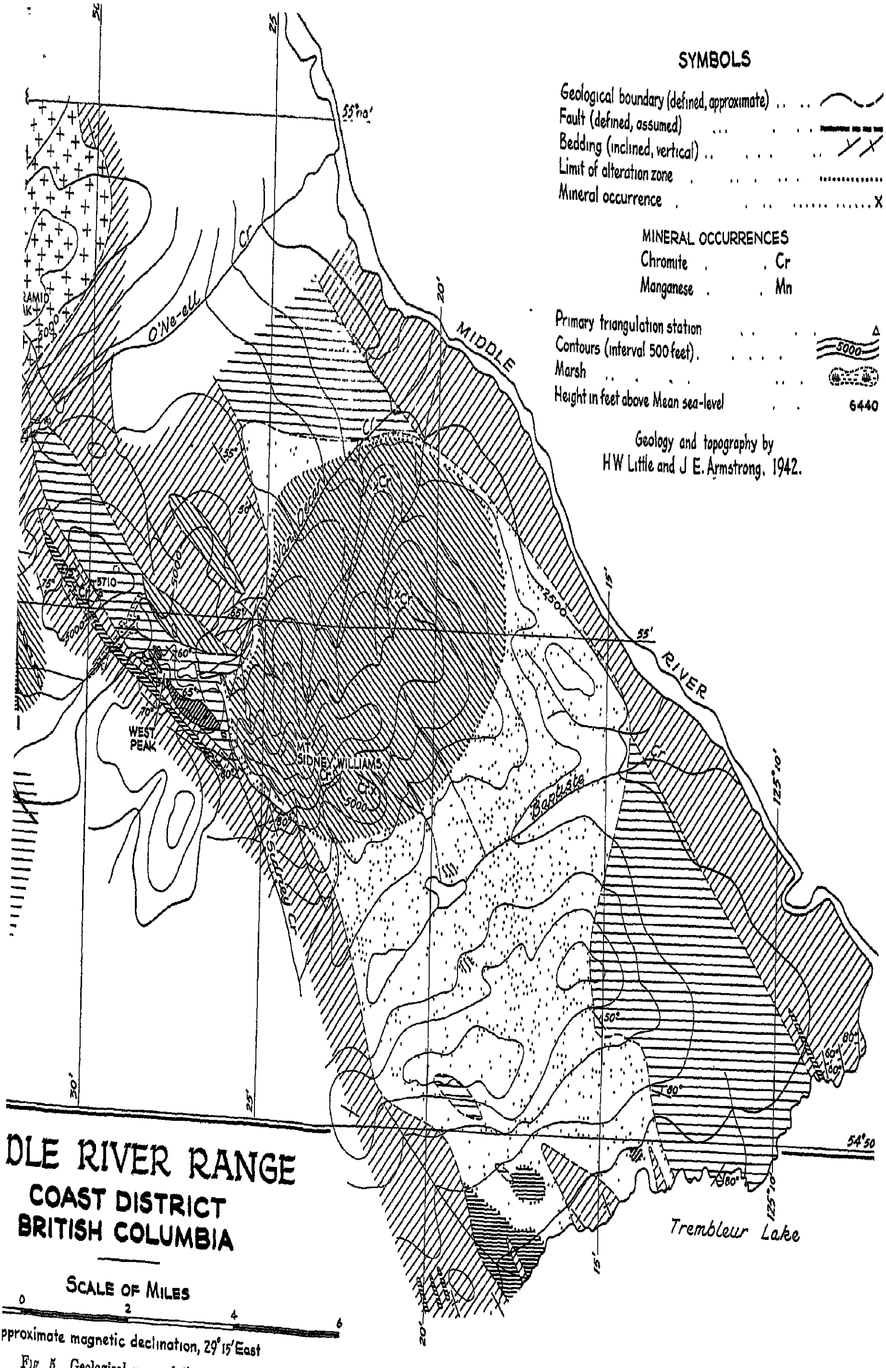
SYMBOLS
Geological boundary (defined, approximate)
Fault (defined, assumed)
Bedding (inclined, vertical)
Limit of alteration zone
Mineral occurrence . . . X
MINERAL OCCURRENCES
Chromite . . Cr
Manganese . . Mn
Primary triangulation station
Contours (interval 500 feet)
Marsh
Height in feet above Mean sea-level . . 6440
Geology and topography by
H W Little and J E. Armstrong, 1942.
MIDDLE
RIVER
O'Ne-ell
Baptiste
Sidney Cr
MT SIDNEY WILLIAMS
WEST PEAK
Trembleur Lake
55°00'
55'
54°50
125°10'
DLE RIVER RANGE
COAST DISTRICT
BRITISH COLUMBIA
SCALE OF MILES
0 2 4 6
pproximate magnetic declination, 29°15' East

Fig. 5 Geological map of the area

Zobel, and these are reproduced in figure 4. While the ultrabasic batholith is somewhat larger in dimensions, the effect would be similar.

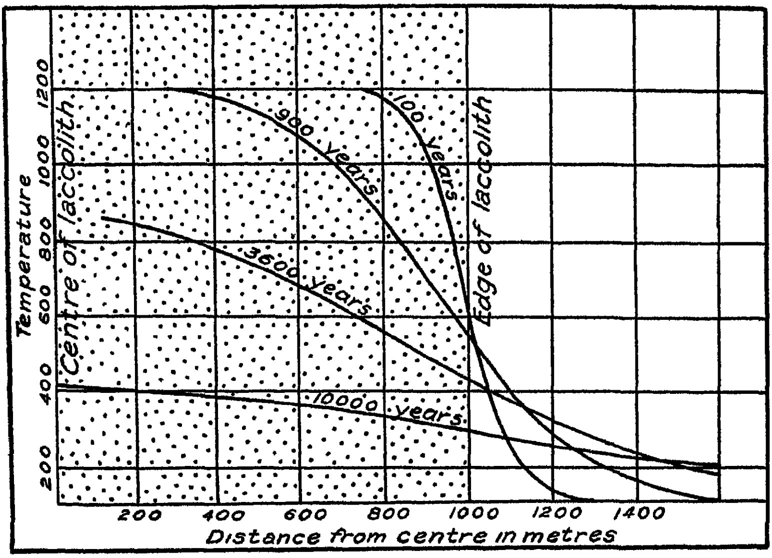

Fig 4 Cooling curves of a spherical or lenticular igneous body intruded at 1200° C. into shales at 100° C. Radius of body is 1000 metres (After Ingersoll and Zobel.)

The writer proposes that the ultrabasic magma was intruded at about 1200° C. into sediments that were comparatively rich in water. As the sediments reached their maximum temperature, hot aqueous solutions derived from them and containing alkaline silicates and carbonates, migrated into the cooling magma. At the chilled borders, where the temperature was much less than in the interior, the reaction favoured was that of conversion of the ultrabasic rock, and the sediments immediately adjacent, to talc and carbonate. The alkaline silicate solutions, which now contained little carbonate, migrated into the main mass of ultrabasic rock, which was at a higher temperature, and brought about serpentinization of olivine and pyroxene.

The preceding calculations cannot be applied to xenoliths, which must have been heated almost to the temperature of the enclosing rock, as heat would have flowed into them from all sides. Argillaceous xenoliths have been baked to an exceedingly hard, fine-grained aphanitic mass. The results should be similar, except for the effects of pressure, to heating the rock in an open crucible, because the ultrabasic rocks would act as an absorbent of aqueous solutions. Where calcium and magnesium are present, tremolite is likely to be formed, but in the later stages of cooling retrograde effects may disguise the earlier high temperature effects. If potassium is present, sericite may be formed.

In one large xenolith a bed of limestone 5 feet thick has been wholly recrystallized to coarse calcite containing euhedral crystals of ilmenite. Considerable chlorite was also introduced into the rock. A sample of calcite was ground to —100 +200 mesh and heated in an electric furnace until the liquid inclusions it contained exploded. The temperature at which the explosions began was 295° C. at one atmosphere pressure. This corresponds to a temperature of 385° C. if the depth at which crystallization took place is assumed to have been 3 miles (Beland, 1946). A sample of ilmenite treated in the same manner indicated a temperature of crystallization only 2° lower. This temperature of 385° indicates the lowest temperature at which hydrothermal solutions deposited calcite and ilmenite. It is apparent that hydrothermal solutions were active in the ultrabasic rock long after the body itself began to cool.

The Origin of the Ultrabasic Rocks. A theory that would account for the origin of the ultrabasic rocks of the Middle River Range must satisfactorily explain the following features:

1. Sills composed almost entirely of olivine have parallel walls and tremendous length, indicating a high degree of fluidity for the magma that emplaced them. On the other hand, bodies consisting of peridotite or norite-pyroxenite are lenticular or oval in shape.

2 In the broad sill-like bodies of ultrabasic rocks, differentiation into zones of peridotite and dunite that are parallel to the walls of the sills has taken place before consolidation.

3. In the ultrabasic batholith dunite forms irregular bodies within the peridotite.

4. In some localities, dunite masses exhibit cross-cutting relationships to the peridotite.

5. Irregular, branching dykes of extremely coarse-grained pyroxenite and pyroxene-rich peridotite intersect the peridotite.

6. There is a distinct break in the petrographic sequence

of rocks from ultrabasic to acidic types; there are no representatives in the area of those types which lie between the norite-pyroxenite complex and soda granite. There is apparently also a time-gap of from 45 million to 140 million years between the time at which the ultrabasic and acidic rocks were emplaced, the latter having been injected after the former.

This complete lack of gradational phases between the two extreme groups of igneous rocks in the area, together with the relatively high proportion of ultrabasic rocks to acidic rocks precludes the acceptance of Bowen's theory (1928, pp. 133-174) of crystal settling from a basaltic magma to account for the origin for the ultrabasic rocks. The writer is forced to the conclusion that the ultrabasic rocks in the Middle River Range were derived from an ultrabasic magma whose composition closely approached that of the rocks derived from it

The first invasion by this magma produced sills at a time when the sedimentary rocks intruded were still flat-lying, or nearly so. The magma was still in a highly fluid state. Sills of dunite were apparently formed at this stage by "filter-pressing". The interstitial liquid migrated farther than the olivine crystals, either during the initial intrusion or shortly afterwards, before the formation of pyroxene crystals could take place. Relatively small, lenticular pockets of the interstitial liquid did remain, however, and these produced lenses of hypersthene and augite.

Narrow bands of alternate olivine-rich and normal peridotite, 2 inches and 1 foot thick respectively, have been described on page 8. A theory that will satisfy the conditions noted in the field is that of periodic release of volatiles. As cooling proceeded, crystals of olivine would settle to the solidified part of the intrusive from the still fluid central part. A sudden release of volatiles would increase the melting point of the liquid phase, so that a layer of peridotite will suddenly solidify, trapping the olivine-rich layer at the base. Periodic repetition of this process would bring about the type of pseudostratification that has been described.

Just prior to the complete solidification of the thickest sills, orogenic movements were initiated. This brought about bending and minor fracturing of the semi-plastic pseudostratification. At the same time fluid material was injected from the hotter parts of the sills which would have been enriched in volatiles owing to the crystallization of most of the silicates. Such magma would contain a large percentage of olivine crystals which would be filter-pressed to form lenticular masses of dunite cutting the peridotite. The interstitial liquid that had been filter-pressed out would migrate farther and form the coarse-grained pyroxene-rich peridotite dykes that have been described. That such a mechanism is plausible is seen in the intrusive mass of dunite described on page 9. In plate 1, fig. 1 it may be seen that the bulbous mass is tubular, and movement of this residual magma took place in a direction perpendicular to the face of the outcrop. During the movement of this magma, olivine crystals were deposited on the walls of the tube and in the apophysis that runs out from it. Later, as the magma cooled, pyroxene crystals were precipitated and these formed the central core. These pyroxene crystals are pegmatitic owing to the relative increase in the volatile content of the last part of the residual magma.

At or near the conclusion of the folding there occurred the final intrusion of magma from the ultrabasic reservoir. This magma was apparently more viscous than that which formed the sills. The intrusive bodies formed at this stage are lenticular or oval in shape, ranging in diameter from 30 feet to 7 or 8 miles. They consist predominantly of peridotite, although the largest bodies contain irregular masses of dunite within the peridotite. Regular banding of these peridotites is common.

These bodies must have been emplaced by thrusting aside the invaded rocks. They could not have been emplaced by stoping, because the ultrabasic magma had a greater specific gravity than the invaded rocks Assimilation of the country rock could have taken place only on a minor scale, if at all, as shown by the presence of numerous xenoliths that have been little affected.

When the ultrabasic rocks were almost completely consolidated there must have remained a final liquid of relatively minor amount, which was injected to form a lenticular stock of norite-pyroxenite complex. Another mass about 1000 feet in diameter occurs within the ultrabasic batholith near Trembleur Lake. The relationship between this mass and the batholith is not known, since it is completely surrounded by steatitized rocks. A comparatively few dykes of norite were injected into the ultrabasic batholith perhaps at the same time. This noritic material may have been present originally in the ultrabasic magma, or it may have been the result of assimilation of basalt or greenstone by the ultrabasic magma just prior to, or during injection.

The deuteric effects that took place during the final stages of cooling of the bodies have already been discussed at length under the heading of serpentinization.

Age of the Ultrabasic Rocks. All of the bodies of ultrabasic rocks have been injected into the Middle Permian succession of sedimentary and volcanic rocks. It has been shown that the intrusion of this magma began while the sedimentary rocks were still flat-lying, or nearly so, so that the formation of the igneous bodies probably took place not long after Middle Permian time.

The Takla series, the lowest part of which is probably Upper Triassic in age, has not been invaded by the ultrabasic rocks, either in the area under discussion or in the surrounding district. In the vicinity of Pinchi Lake, Armstrong (personal communication) found pebbles of serpentine and grains of chromite in a conglomerate of Upper Triassic age.

Ultrabasic rocks are traversed by dykes which are related to granite which intrudes the Takla series, so that the former are definitely pre-granite in age.

The age of the ultrabasics, then, has been established as post-Middle Permian and pre-Upper Triassic, the most likely time being about the close of the Permian, since they had become exposed to erosion in Upper Triassic time.

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DEFINITIONS OF ARKOSE

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ABSTRACT. An examination of the definitions of the term *arkose* in the current literature shows that the rock is described by different writers as 1) a sandstone with a considerable amount of feldspar, 2) a coarse sandstone or fine conglomerate of quartz and feldspar, and 3) a feldspathic sandstone which looks like a granite. In defining the term in 1823, Alexandre Brongniart stressed the high feldspar content as the distinguishing feature of the rock. Thus, definition 1) most nearly agrees with Brongniart's, whereas the other two limit *arkose* to rocks which Brongniart described as varieties of arkose.

WHILE engaged in the preparation of a report on the geology of the Hot Springs area, Madison County, N.C.,[2] the writer found it necessary to reexamine definitions of the term *arkose*. In view of the revived interest in the classification of sedimentary rocks during the last few years, the results of this brief literature study are presented here.

At least three definitions of the term *arkose* are in current usage. In the definition suggested by the Committee on Sedimentation (Allen, 1936, p. 44), arkose is described as "A sandstone containing 25 or more per cent of feldspars usually derived from the disintegration of acid igneous rocks of granitoid texture."[3] Cordier (1868, p. 218) limits the feldspar composition to 10 to 20 per cent of the rock. Krynine (1940, p. 50) sets 30 per cent as the lower limit, but shows the "average" feldspar composition of his "arkose series" as 25 per cent in a later published figure (1948, fig. 11, p. 151). Whatever the proportion used as the lower limit, many geologists (Barton, 1916, p. 418; Coquand, 1857, pp. 236-237; Cotta, 1855, p. 214; Dana, 1863, p. 83; Grout, 1932, p. 275; Hatch, Rastall and Black, 1938, p. 84; Loewinson-Lessing, 1893, p. 18; Milner, 1940, p. 369; d'Orbigny, 1845, p. 140; Pettijohn, 1949, table 55 and fig. 66, p. 227; Shrock, 1948, p. 122; Twenhofel, 1932, p. 229; and Zirkel, 1866, vol. II, p. 526) have accepted the definition of an arkose as a sandstone containing a substantial amount of feldspar and derived from granitic rocks.

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[1] The writer is grateful to Messrs. Adolph Knopf, John Rodgers, and Gerald V. Carroll for helpful suggestions.

[2] To be published as a bulletin of the North Carolina Department of Conservation and Development.

[3] Sandstone is defined as a consolidated rock composed of cemented sand grains between 1/16 mm. and 2 mm. in diameter (Allen, 1936, p. 46).

Cayeux (1929) defines arkoses as "very coarse feldspathic or kaolinitic sandstones derived from granitic or gneissic sand" (translated from p. 207). "Arkoses form a transition, by mineral size, between the finer components of pudding-stone [lower limit: 5 mm., pp. 9, 119-120] and sandstone, properly speaking ["5 mm. to 0.5 mm.," p. 124], but with a greater affinity to sandstone than to pudding-stone" (p. 214). Thus the distinction made by Cayeux between feldspathic sandstone and arkose is one of size, not of total feldspar content. Bertrand (1928, p. 448), Dana (1894, p. 82), and Holmes (1920, pp. 36, 282) also limit *arkose* to rocks intermediate in grain-size between psammites and psephites. This definition is supported by that of Omalius d'Halloy (1831, p. 537) which states, in translation, that an arkose is a "conglomeratic rock composed of quartz and feldspar."

De Lapparent (1906) considers arkoses as "rocks formed principally of quartz and feldspar, perhaps with mica, in a manner simulating a true granite, whence the name *decomposed* or *regenerated granite* is given to them" (translated from p. 685). Thus, in addition to requiring that the rock contain a considerable amount of feldspar, de Lapparent and others (Krynine, 1940, p. 50; Murchison, 1867, p. 347; Pirsson and Knopf, 1947, p. 251; and Rosenbusch, 1898, p. 393) impose or suggest the second condition that arkose simulate a granite in appearance.

In view of this divergence in the current definitions of *arkose*, the present writer turned to the original definition of the term by Alexandre Brongniart (1823, pp. 497-498). The term was introduced by Brongniart in an attempt to limit the use of the word sandstone (*grés*), which he felt included too many diverse rocks. He retained the word sandstone for those rocks consisting almost entirely of granular quartz (as distinguished from glassy quartz in quartzites). His definition of the term, literally translated, is, "Arkoses are composed of large grains of glassy quartz and of feldspar, mixed together unequally and including as fortuitous constituents mica, clay, often kaolin, etc."

Why Brongniart chose the word *arkose* is not known. Insofar as the present writer has been able to determine, the word did not exist before 1823. In his original definition of the term, Brongniart (1823, p. 498) does not indicate the derivation of the term but does say that it is one of the names he has adopted at the suggestion of Omalius d'Halloy. The derivation of the term is not indicated in those publications by Brongniart, Omalius d'Halloy, and other writers available to the writer.

The present writer tentatively suggests that the term *arkose* was taken from the Greek *archaios* meaning ancient or primitive. The points which bear on this view are as follows:

1) Brongniart states (1826, p. 30) that rock names should be derived, with few exceptions, from the "sonorous" Latin and Greek languages and should be easily pronounced by all peoples.

2) Brongniart states (1826, pp. 123-124) that the majority of arkoses studied by him rest directly on granite and indicate ". . . a quite ancient epoch of formation" However, elsewhere in the same paper, Brongniart concludes that there are younger arkoses and age is not implicit in the definition of the term.

3) Brongniart wrote his papers with clarity and in detail. This suggests that he omitted the derivation of the term as superfluous, in view of its similarity to the Greek *archaios.*

4) The use of *k*, rather than *ch* or *c*, in spelling the word is rather curious. Brongniart may have used *k* in preference to the others to insure that the term would be pronounced similarly in French, German, and English, or he may simply have followed the example of Jurine (1806, pp. 373-374), who had previously proposed another rock name, *arkesine*,[4] from the same Greek root.

Brongniart discussed the arkoses more fully in a 51-page paper in 1826. Inasmuch as no statement of his views appears in the modern literature (probably because his publications are not easily available) and as it has a bearing on the present problem, excerpts from his petrographic discussion are translated below. (His views on genesis of the rock, though interesting, do not bear on the problem here.)

> Arkose is a rock of grainy texture formed principally as a result of mechanical aggregation.
>
> It is *composed essentially* of large grains of glassy quartz and grains of lamellar, compact or clayey feldspar: these two minerals are often mixed in more or less equal quantities but more often quartz is dominant. It includes, as *accessory constituents*, mica,

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[4] Proposed for the talc- and chlorite-bearing hornblende granite on Mont Blanc, the term is no longer in use.

lithomarge clay, and kaolin, in quantities less than either quartz or feldspar.

The *fortuitous ingredients* (*parties accidentalles*) which are disseminated through the arkose are many [and a list of those minerals which he identified is presented].

This rock has no distinct small scale structure, and rarely even on a large scale, and it is generally thick-bedded.

Its texture is essentially grainy; grains angular, ranging from millet-seed size to pea size. The mass of the rock was evidently formed as a result of mechanical aggregation; the irregular and angular form of the grains and especially their "limitation," such that they do not penetrate one another, is the proof of it . . .

The arkoses, sometimes so distinctly characteristic that one cannot confuse them with any other rock, in some cases present vague, uncertain, or incomplete characteristics.

When the feldspar content is low and the quartz content high, the rock passes into quartzite, if the quartz is glassy, or into sandstone (grés) if the quartz is finely granular .

The varieties which are present in these rocks are not numerous and may be reduced to the following:

Common arkose

> Composed of glassy quartz and grains of feldspar with very little mica, quartz dominant. Color grayish or whitish . . .

Granitoid arkose

> Grains of quartz, of lamellar feldspar and of mica, more or less disposed as in granite, feldspar dominant.
>
> This rock does not differ from granite except that it is evidently formed by aggregation.

Miliary arkose

> Grains of quartz and feldspar, as large as millet seeds at the most; disseminated colored clay; quartz dominant; little mica . . . does not differ from common and granitoid arkose except by grain size . . .

It is evident from the excerpts quoted here that Brongniart considered the granitic appearing rocks as one of several forms of arkose. Therefore it appears that the second condition imposed by the de Lapparent definition, i.e. that the rock simulate a granite in appearance, was not considered necessary by Brongniart. Furthermore, the subjectivity of the condition makes it undesirable in modern usage.

Although Brongniart specified "large grains" in his original definition, it is the present writer's belief that it is not an essential part of his definition. This view is supported first by his making arkose a subdivision of sandstone (1827, p. 43) and second by his defining fine-grained or miliary arkose as one of the varieties of arkose. Therefore, and because the size connotations stressed by Cayeux have not been generally adopted in most countries, it seems unlikely that Cayeux's definition will prove acceptable to most geologists, despite the support it receives from that of Omalius d'Halloy.

Thus, of the three definitions of arkose in current usage, that of the Committee on Sedimentation appears to reflect the meaning intended by Brongniart best and does so in quantitative terms and in conformity with the usage of many geologists. The other two definitions may well be retained for varieties of arkose, in keeping with the procedure of Brongniart himself.

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REVIEWS

Two Problems of Marine Geology: Atolls and Canyons; by PH. H. KUENEN, Pp. 69; 4 pls. Kon. Ned Akad. Wet., Verh. (Tweede Sectie), Dl. XLIII, No. 3, 1947.

I *The Borings of Maratoea Atoll and the Coral Reef Theory of Glacially Controlled Subsidence.*

An important contribution to the study of coral reefs is furnished in the form of quantitative data concerning depth of coral limestone. Two borings were made by the Bataafsche Petroleum Maatschappij on the raised atoll of Maratoea, about 80 km. northeast of Borneo. One hole passed through 261 m. and another through 429 m. of coral limestone and related deposits. The cores are interpreted as indicating a reef margin where coral is interfingering with coral detritus and lagoonal deposits.

After reviewing arguments against theories of atoll formation on stationary platforms, Kuenen suggests that the great thickness of coral indicates upward coral growth during subsidence of an underlying platform. During the Pleistocene lower temperatures halted coral growth in marginal areas of the coral seas and mechanical wave abrasion soon reduced the reefs to platforms graded to the lower sea level. Kuenen suggests that in the central coral seas solution by temporarily undersaturated sea water lowered the elevated reefs while corals continued to grow at the lower sea level. With the subsequent rise of sea level corals grew up from the platforms.

Although this theory of a multiple origin of coral reefs is not new, the subsurface evidence now available gives it much greater importance However, subsidence at Maratoea may have been local; only additional data will indicate its wider importance. Kuenen's choice of a name for this theory is unfortunate for the subsidence is not glacially controlled as the term indicates.

II *Submarine Canyons and Their Formation.*

Kuenen outlines some known facts concerning submarine canyons and critically reviews seven principal hypotheses of origin. He favors canyon excavation by currents of suspended sediment moving down the continental slope. During the Pleistocene, with lower sea level, large quantities of loose sediment were exposed along coasts which favored development of heavy suspensions by storm waves. Slumping and sliding would be important both in forming the initial furrow and in widening canyons.

Kuenen found experimentally that a heavy suspension of sand released on an underwater slope of 13° attained a speed of 30 cm/sec. He estimates that a velocity of 1 to 4 m/sec might be expected under natural conditions. At the present it seems to be but a matter of opinion whether a suspension current of this velocity could erode any but the most loosely consolidated sediments to the extent of the known canyons. DWIGHT R. CRANDELL

Gesteine und Minerallagerstatten. Erster Band. Allgemeine Lehre von den Gesteinen und Minerallagerstatten; by PAUL NIGGLI, with special assistance on Part IV by ERNST NIGGLI. Pp. 540, 335 figs. Basel, 1948 (Verlag Birkhäuser, 42 Swiss francs in paper, 46 Swiss francs in cloth).—This is the first in a projected series, now approaching completion, of three volumes on rocks and mineral deposits. The present volume has been designated as Number 3 in the Mineralogical-Geotechnical Series of Textbooks and Monographs in the Realm of the Exact Sciences, and is concerned solely with some of the fundamental principles of petrology and the science of mineral deposits In the foreword, the author states that he has found the customary division into igneous, sedimentary, and metamorphic rocks and mineral deposits to be very unsatisfactory, especially in view of the usual lack of preliminary discussion of the basic principles underlying their formation. His chief objection to the three-fold division is based on the resulting development of genetic concepts, which are completely divorced from each other, for each of the three classes of rocks This volume, therefore, is an attempt to bring these principles together in a new way Detailed descriptions of rocks and their associated mineral deposits will appear in the two forthcoming volumes entitled "Die exogenen Gesteine und Minerallagerstatten" and "Die endogenen Gesteine und Minerallagerstatten" respectively.

Part I is concerned with the principles of geochemistry and crystal chemistry, and includes descriptions of the most important rock-forming minerals and ore minerals with special emphasis on their atomic structure. Of particular note are the fine photographs of models illustrating the atomic structure of minerals.

Part II, entitled "Das Gefuge," is a detailed treatment of the structure and texture of rocks and ore minerals from both the macroscopic and the microscopic viewpoint. Sections on petrofabric analysis, the application of statistical theory to petrologic problems, and the methods of representing graphically and numerically the mineral and chemical composition of rocks are included Zoning in crystals, intergrowths, overgrowths, porosity, permeability, corona structures, and the size and shape of grains also are treated. These pages are profusely illustrated with excellent line drawings of a great variety of fabrics.

Part III is a treatise on the physical chemistry of the genesis of rocks and ore deposits. Following the theoretical discussions is a separate section on the important processes in mineral formation.

Part IV discusses the physics of the Earth's outer shell. Numerous tables present the fundamental physical constants of rocks and minerals. A section on applied geophysics describes eight current geophysical methods used in the search for mineral deposits and subsurface structural data.

Part V is a summary of the author's classification of rocks and mineral deposits which will appear in greater detail in volumes II and III of this series. In brief, all rocks and mineral deposits are divided into two major groups, the exogenous and the endogenous. The exogenous rocks and mineral deposits are formed in the border zones of the lithosphere, atmosphere, and hydrosphere. All the sedimentary rocks and all weathering products are included in this category. The endogenous rocks and mineral deposits are those that are formed within the lithosphere. They are subdivided into katathermal rocks and mineral deposits and epi- to mesothermal rocks and mineral deposits depending on the intensity conditions under which they were formed. The katathermal division includes the igneous rocks, high temperature veins, ortho- and paragneisses, migmatites, and the products of high temperature hydrothermal and pneumatolytic alteration. The epi- to mesothermal division includes the rest of the metamorphic rocks, low temperature veins, and the products of low temperature hydrothermal alteration.

There are no references to the periodical literature except in Part IV, Section C, on applied geophysics, but the book contains an extensive list of current reference works.

To the beginning student this mode of presentation will probably prove to be confusing. There is no mention of any connection between particular rock-forming minerals and specific geologic environments. A great many fabrics are carefully described and superbly illustrated, but the genetic implications of the fabrics and the close relationship of many fabrics to specific rock types are not brought out. As a result, the student is faced with a bewildering array of mineral and fabric names that conveys little of geologic significance.

On the other hand, the advanced student will find a great wealth of pertinent information assembled in one volume. Professor Niggli has performed a notable service in presenting the principles of statistics, physical chemistry, and crystal chemistry from the geologic viewpoint.

CHARLES B. SCLAR

ERRATA

In the paper by Donald V. Higgs, August number, p. 582, the cut, figure 5, should have been inverted. P. 583, second reference, should read "Bull. Geol. Soc. America, vol. 60, no. 4."

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A PRE-WISCONSIN FOREST SUCCESSION IN THE PUGET LOWLAND, WASHINGTON[1]

HENRY P. HANSEN AND J. HOOVER MACKIN

Presented before the Ecological Society of America, AAAS Meeting, Vancouver, B.C., June 1949

ABSTRACT. Pollen profiles from existing peat bogs in the Puget Lowland provide a clear picture of forest successions and climatic changes from the time of retreat of Wisconsin ice to the present. This paper applies the pollen analysis technique in a study of compacted peat layers and peaty clays that make up part of one stratigraphic unit in a complex sequence of pre-Wisconsin Pleistocene deposits that underlie the central part of the Lowland. While there are some nondepositional or erosional gaps, the forest succession as reflected in the pollen profiles indicates that the unit represents an interglacial age, including a phase of ice recession, a climatic optimum comparable with the present, and the readvance of ice of the next glacial age. These conclusions are based on samples collected from only one composite section and are therefore tentative, but the results indicate that pollen analysis will be a valuable tool in working out the pre-Wisconsin Pleistocene stratigraphy in the Puget Lowland

INTRODUCTION

THE principal topographic elements of the central part of the Puget Lowland are north-south trending drumloidal hills and intervening troughs occupied by the branches of Puget Sound. The drumloids are veneered by the Vashon (Wisconsin) till sheet; their cores consist of flat-bedded gravels, sands, and clays, with intercalated layers of peat and lenses of till, comprising the "Admiralty sediments." Bretz (1913), following Willis (1898) in part, concluded from these and other relationships (1) that the Admiralty sediments were deposited chiefly during recession of the Admiralty glacier, (2) that the aggradational plain so formed was trenched by streams during an interglacial age, and (3) that this interglacial hill and valley topography was modified into the present trough-

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[1] The first author is grateful to the John Simon Guggenheim Memorial Foundation for a fellowship during the year 1947-48, which enabled him to devote his time to research, and to the American Philosophical Society, and the General Research Council of Oregon State College for grants to defray laboratory and field expenses.

drumloid topography by the Vashon glacier during Wisconsin time.

A general restudy of the glacial geology of the Puget area now in progress (Mackin, 1941) indicates that at least two pre-Wisconsin drift sheets are separable in southern drift border areas on the basis of weathering contrasts and other conventional mapping criteria. The Admiralty sediments that underlie the Vashon till in the central part of the Lowland also include deposits of more than one pre-Wisconsin age, but the conventional methods are not applicable because such weathered mantles as may have formed during interglacials have been destroyed in process of emplacement of later till sheets; subdivision of the Admiralty sediments is essentially a stratigraphic problem with no good basis for assigning time values to the disconformities that occur within the sequence. A closely related and equally serious difficulty has been a lack of evidence as to the climatic conditions that prevailed during the deposition of Admiralty lithologic units other than till sheets, that is, to whether a given sedimentary unit was deposited in immediate proximity to glaciers, or during a period of advance or retreat, or during a warm interglacial.

Pollen profiles of sections from existing peat bogs throughout the Puget Lowland provide a clear picture of the major trends of forest succession during late glacial and recent time (Hansen, 1947a). In this postglacial sequence, the base of the peat marks the time of withdrawal of the ice, and the top of the living bog is the present, so the duration of the period of accumulation of the peat section is therefore reasonably well known. Forest succession indicated by the pollen profiles can be correlated with other kinds of data on a regional scale and furnish an excellent record of postglacial climatic trends and the many other factors, such as fire and varying physiographic conditions from place to place, that influence the normal forest development from pioneering types to climax types.

It is evident that the pollen analysis technique, applied to the compacted peats and peaty clays that occur in the Admiralty sediments, may furnish answers to some of the pre-Wisconsin glacio-stratigraphic questions outlined above, and that the mass of data bearing on postglacial forest development provides a sound basis for interpreting pre-Wisconsin forest successions in the same area. This paper illustrates the use of pollen counts in a study of a composite section of an Admiralty stratigraphic unit that apparently represents an interglacial age, including a phase of ice recession, a climatic optimum more or less comparable to the present, and the readvance of the ice of the next glacial stage. Microscopic work and inferences as to forest types, climatic conditions and duration of the interglacial are by Hansen. Discussion of field relations, by Mackin, is confined to the sections sampled in the course of this joint study.

DESCRIPTION OF THE EVERETT GORGE SECTION

The Everett Gorge Section is exposed in the sides of a narrow "inner valley" cut within the last 25 years by waste water from an Everett reservoir (SW¼, Sec. 30, T. 29 N., R. 5 E.; see Everett quadrangle). The Gorge is essentially vertical walled, about 1,000 feet long, and with a maximum depth of about 125 feet. Sets of samples A to D (fig. 1) were taken with the aid of ropes and ladders from different parts of the walls, such that the top of a lower unit sampled downstream (as A), is traceable by eye to the base of the next higher unit (as B) sampled upstream. Because pollen grains are not well preserved in oxidized sand and gravel, only the silt and clay layers were sampled. Each sample represents a segment of a continuous channel picked into the cliff face far enough to eliminate the possibility of contamination by surface wash. Figure 1 shows the succession of beds and the order in which the samples are numbered.

All of the coarser sediments are strongly oxidized; open-textured gravels are deep red-brown in color and sands are tan and brown. Silts and clays are generally dark gray with shades of brown due to the presence of organic matter rather than to oxidization of iron. Oxidation effects are uniform throughout the thickness of the gravel layers and penetrate a few inches to a foot into overlying and underlying silty clays, these relations suggesting pre-gorge conditions when the gravels carried ground water under pressure in contrast to present conditions under which seepage is usually confined to the lower few inches of each gravel layer. The significant point is that oxidation varies directly with permeability and is clearly due to circulation of subsurface water; the alternation of oxidized sand-gravel layers and unoxidized silt-clay layers does not indicate lapses of time during deposition of the sequence.

The lower part of the section (units A and B) consists predominantly of dark gray and brown silty clays with which are interbedded sheets and lenses of well sorted gravel five to ten feet in thickness, with average larger pebble sizes to two inches. Thick layers of cross-bedded sand are notably absent. Interbedding of the gravel and the silty clays certainly does not mean changes in depth of water or other regional conditions of deposition. The silty clays are overbank deposits, formed in floodplain lakes and swamps, and represent the finer fractions of the suspended loads of through-flowing streams. The gravels are the channel deposits of the same streams, and represent the coarsest fractions of their bed loads.

The general picture suggested by the lower part of the Everett Gorge section is that of very slow aggradation by meandering streams flanked by floodplain lakes and swamps in which the silty clays and peats were laid down. The fact that this part of the section consists dominantly of overbank materials, and only in small part of channel gravels, indicates that the rate of lateral shifting of the streams was slow relative to the overall rate of upbuilding of the aggradational plain (Mackin, 1948, p. 502-3). It is important, in this connection, to note that the gravel lenses occupy channels cut in the finer-textured sediments, and that there is no way to determine the thicknesses of silty clay and peat that were scoured away by laterally shifting meanders during the period of deposition. There is, in other words, no geologic basis for assigning time values to the discomformities at the base of each channel gravel sheet, and it is therefore evident that estimates of rates of deposition of the silty clay and peat now seen in gorge walls can provide only a minimum figure for the duration of the period of time represented by the sequence.

The upper part of the section, particularly above C 53, is composed largely of coarse, cross-bedded sand with scattered small pebbles, probably representing channel deposits of streams of braided habit (fig. 1). The marked decrease in grain size from the channel gravels in the lower part of the section does not indicate a decrease in size of rock fragments supplied to the streams at their sources; it suggests, rather, that under changed conditions of rapid aggradation, the boulder and pebble fraction of the stream loads were being selectively deposited in aggradational fills farther "upstream," so that only the sand fractions reached the Everett locality (Mackin, 1948, p. 505-6). Similarly, the fact that laminated clays and silts, which predominate below, make up only a small part of the upper deposits is probably due to rapid lateral shifting of the braided channels as contrasted with relatively stable channel conditions that prevailed during accumulation of the lower part of the section.

An indirect line of evidence bears on the reason for the change in type of sedimentation during the deposition of the Everett section. Overbank silts and clays in the transitional part of the sequence (the C unit) are described as "fibrous" in the columnar section because, where they are exposed to the gentle washing action of the present stream, they carry a dark brown "hairy" coating made up of partly oxidized shreds of woody material. Similar woody fluff accumulates on screens used in sizing certain Puget tills deposited by glaciers that were, in the immediate vicinity, advancing through forests. The presence of macerated logs in and under these till sheets indicates that the fluff is, so to speak, the woody equivalent of rock flour. Its occurrence, together with numerous abraded wood slabs and rounded knots, in the silts and clays in the transitional part of the Everett section, and the absence of these materials in the lower part of the section, suggests that the change from stability or very slow aggradation by meandering streams to rapid aggradation by braided streams was due to advance of glaciers into the drainage basins of these streams.

DESCRIPTION OF THE POSSESSION POINT SECTION

It is virtually certain, on the basis of relationships in the Admiralty sediments elsewhere in the Puget Lowland, that the sequence exposed in the Everett Gorge is underlain at some depth by a till sheet marking the base of the major pre-Wisconsin depositional unit of which the Everett Gorge beds are a part, and it is evident that sampling for pollen analysis should, if possible, extend downward to the till. The lowest exposures in the Gorge are about 95 feet above sea level. Clays, silts, and gravels are encountered in pits and hill slope trenches at lower levels, but all of the lower valley sides and wave-cut cliffs in the vicinity are so heavily covered by creep material, landslide debris and/or Vashon till that no continuous section is available (absence of exposures at and just above sea level is

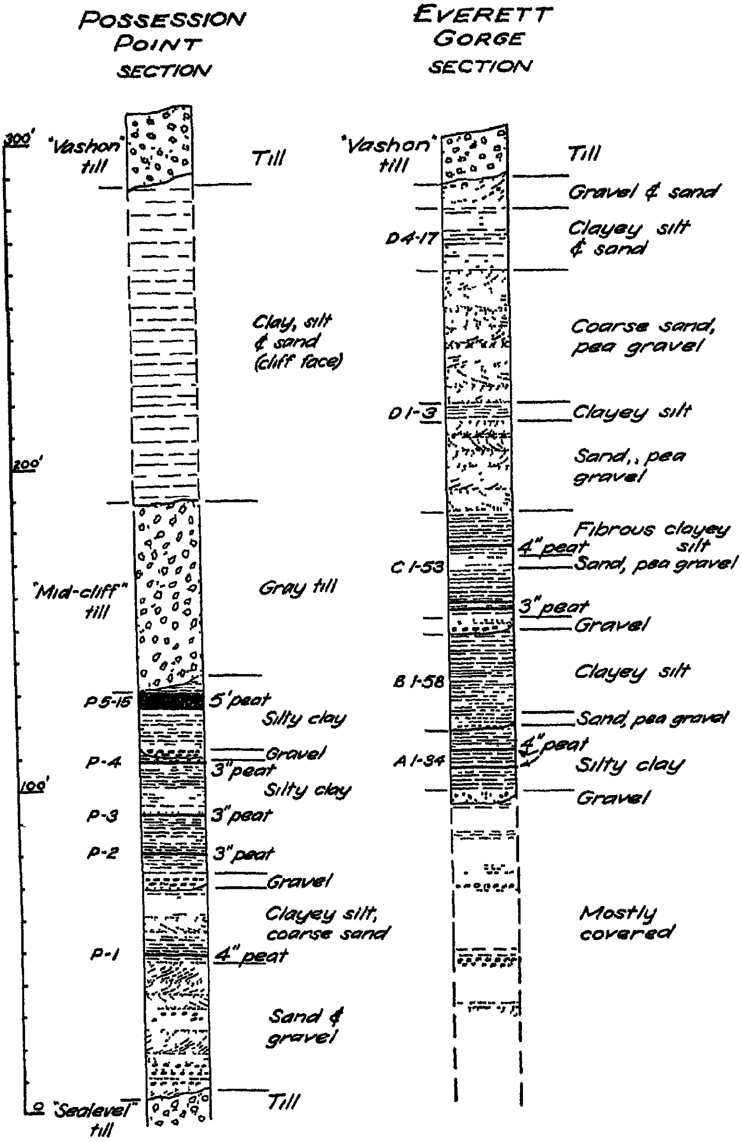

Fig 1. Diagram of Possession Point and Everett sections showing sequence of sediments as discussed by Mackin and interbedded peats and pollen-bearing clays and silts that were analyzed for pollen.

due chiefly to a railway embankment that follows the shoreline and has checked wave erosion along the east side of Puget Sound). The nearest continuous section seen in reconnaissance is at Possession Point, a fresh-wave-cut cliff at the south end of Whidby Island, about nine miles west-southwest of the Everett locality.

The Possession Point exposures show a till at sea level, overlain by poorly sorted sand and gravel with ice shove and slump structures. These outwash materials grade upward between 50 feet and 75 feet above sea level into dark gray and brown peat-bearing silty clays with lenses of stained gravel identical in appearance with the lower part of the Everett Gorge section (fig. 1). Samples P 1 to P 4 represent thin peat layers occurring 50 to 110 feet above sea level; samples P 5 to P 15 were taken from a 5-foot bed of massive peat at 125-136 feet above sea level. The thick peat layer is overlain, with undulating, ice-scoured contact, by dense gray till (hereafter, the "Mid-cliff till") to 180-190 feet, laminated blue clay, silt and sand to about 290 feet, and Vashon till to the top of the cliff exposures at about 329 feet (fig. 1).

Similarity in lithology, and the general correspondence in elevation above sea level, indicate that the peat-bearing sediments below the "Mid-cliff till" at Possession Point probably correlate with and predate in part the lowermost (A and B) sediments in the Everett Gorge section (fig. 1). Exposures of peat-bearing silty clays with stained channel gravels at the same general elevations in sea cliffs and artificial excavations between the two localities tend to confirm the correlation and to indicate that the unit was formerly continuous throughout this portion of the Puget Lowland. The fact that individual beds of peat, clay, and gravel are lenses, nested in cut and fill arrangement, means that precise elevations above sea level has little significance concerning bed-by-bed correlation between the Everett and Possession sections.

GEOLOGIC INTERPRETATION OF THE EVERETT GORGE-POSSESSION POINT SECTION

The composite section based on this tentative correlation is divisible into three parts for purposes of discussion: (1) the "sea-level till" at Possession Point with overlying poorly sorted sand and gravel grading upward at 50-75 feet above sea level into peat-bearing silty clay with well-sorted channel gravel, indicating recession of a glacier from the central part of the Puget Lowland, (2) the peat-bearing silty clays and gravels between about 75 and 130 feet above sea level at Possession Point, and between 95 and 150 feet above sea level in the Everett Gorge, indicating floodplain sedimentation under conditions of physiographic stability, and (3) the coarse, cross-bedded sands between 150 feet and the top of the Everett Gorge section, indicating aggradation by braided streams, probably proglacial. Segments (1) and (3) consist of coarse clastic sediments that could have been deposited within a few tens or hundreds of years. The fine-textured clastic and organic sediments comprising segment (2) doubtless represent a much longer period of time, perhaps of the order of one thousand to several thousand years as a minimum. The point, mentioned earlier, that needs emphasis here is that neither the Everett nor Possession Point section is a continuous depositional sequence, and that there is no geologic basis for estimating the time values of the disconformities that occur at the base of each channel gravel sheet. Similarly, the geologic evidence proves only that the central part of the Puget Lowland was free of ice when the sediments of segment (2) were laid down; there is no basis for conclusion as to climatic conditions. It is therefore uncertain, on the basis of the geologic relations alone, whether the Everett-Possession section represents (a) a major interglacial age, or (b) a minor recession and readvance of ice during a glacial age (Willis, 1898, p. 150; Bretz, 1913, p. 174).

AGE OF THE EVERETT GORGE-POSSESSION POINT SECTION

The position of the Everett Gorge-Possession Point stratigraphic unit in the Pleistocene succession in the Puget Lowland is problematic and is not critical for present purposes, but evidence bearing on this aspect of the case needs to be outlined if only to eliminate a misunderstanding:—

If only the section in the immediate vicinity of the Everett Gorge were considered, the inclination would be to identify the advancing glacier suggested by the cross-bedded sand in the upper part of the Gorge section with the Vashon till that caps the sequence. But the flat-lying sediments at Everett and Possession Point form part of the cores of drumloidal hills that are veneered by Vashon till, not only on the top but also on the flanks, and the till sheet doubtless continues under the Puget trough that separates the two localities. As indicated in the introduction, this relationship supports Bretz' view to the effect that there was probably a considerable interval of time and more or less extensive stream valley cutting between deposition of the sediments and emplacement of the Vashon till that rests unconformably on them. The "Mid-cliff till" and overlying glacio-lacustrine sediment at Possession Point are beneath this unconformity.

Wherever the peat-bearing sediments are being eroded by waves or rivers at the present time, the beach and stream gravels include rounded boulders of compacted peat. The high degree of compaction is believed to be due, not to the thin mantle of sediments that rest on the peat layers, but to the weight of several thousand feet of ice that has covered the area subsequent to the deposition of the peat. It is therefore significant that, wherever the Vashon till sheet rests on peat-bearing phases of the sediments, it includes boulders and subrounded slabs of peat that evidently were, when incorporated in the till, quite as compact as at the present time. The "Mid-cliff till" at Possession Point, on the other hand, is locally discolored and contaminated by the same peaty materials where it rests on them, but the till contains no fragments of compacted peat. The implication is that the "Mid-cliff till" was formed during the glacial age that immediately followed the deposition of the peat-bearing sediments, and that the Vashon till, which overlies the "Mid-cliff till" unconformably, was formed during a glacial age, after the peat beds had been thoroughly compacted. It follows that if the "Mid-cliff till" ever covered the Everett locality, it was removed by erosion previous to emplacement of the Vashon till.

These relations, taken together, indicate that the Everett Gorge-Possession Point peat-bearing sediments could not have been formed during the interglacial immediately preceding the Vashon (Wisconsin) glacial age, but most date from a time earlier in the Pleistocene history of the Puget Lowland.

METHODS AND TECHNIC

Samples were taken from the exposures at 6-inch intervals from the clay and silt strata and other inorganic layers that seemed promising for pollen analysis. At least one sample was taken from the thin peat laminae, and if the peat layer was 3 or more inches thick, several samples were taken. Much of the clay-silt strata was brownish in color due to the amount of organic material present. Samples for pollen analysis were made up for foot intervals in the clay-silt layers, but in the peat and more organic-bearing strata, all samples were prepared and analyzed. Samples were taken at 6-inch intervals in the upper Possession Point peat stratum and all of them were analyzed for pollen. Samples in the D series in the upper part of the Everett section were also prepared and studied, but few pollen grains were present, and its pollen profiles are omitted. Likewise, levels 1 to 10 and 13 to 26 in the A series were found to have insufficient pollen to use for interpretation. The pollen profiles for the B and C series, however, represent largely consecutive 1-foot horizons with the exception of a few at closer intervals in the thicker peat layers. The greatest intervals between the pollen profile horizons occur in the lower Possession Point strata, where they may be as much as 30 feet. Actually the pollen profiles represent some 150 feet of sediments (fig. 1).

In the preparation of the sediments for pollen analysis, those containing a high fraction of inorganic matter were treated with commercial hydrofluoric acid to remove at least the siliceous elements. This tended to concentrate the limited number of pollen grains where necessary, and also made identification easier. Samples of the pollen-bearing silts and clays were first agitated vigorously in water, then washed and decanted several times to remove the coarser particles. About 4 cc. of the remaining residue was immersed in hydrofluoric acid to a depth of 2 inches in a paraffin-coated paper cup. The samples were left in the acid for several days, the length of time depending upon the amount of silt present, and then washed until all traces of the acid were removed. The remaining residue, now brown as a result of the concentration of the organic material, was boiled for 10 minutes in a weak potassium hydrate solution with a few drops of gentian violet stain. If all of the acid was not removed, the gentian violet turned green and did not seem to stain the pollen grains. The residue was then washed through a fine-mesh tea strainer, centrifuged, and mounted in warm glycerin jelly. The brown peat was not treated with hydrofluoric acid, but a stronger solution of potassium hydrate was used and it was necessary to boil the sample for at least 20 minutes in order to effect deflocculation. An enormous amount of pollen was present in the peat strata, probably as a result of the compaction caused by the overlying sediments. The weight of the overlying sediments, however, apparently caused many of the pollen grains to be crushed and broken, which may have been augmented by the rigorous treatment in the preparation of the samples for microscopic analysis.

One hundred or more pollen grains were identified from each level represented in the pollen diagrams. Pollen of lodgepole pine (*Pinus contorta*) and western white pine (*P. monticola*) were separated by the size-range method (Hansen, 1947a). Mountain hemlock (*Tsuga mertensiana*) pollen was readily separated from that of western hemlock (*T. heterophylla*) by the presence of air bladders on the former. Pollen of Douglas fir (*Pseudotsuga taxifolia*) is distinct from that of any other forest tree in the Pacific Northwest and offers no difficulty in its identification, even when broken. An attempt was made to separate the species of *Abies* by the size range method, but as there is considerable overlap in their size ranges, and so many of the grains were broken, only subalpine fir (*A. lasiocarpa*) was separated with a feeling of any degree of accuracy. The other fir pollen could conceivably represent grand fir (*A. grandis*), silver fir (*A. amabilis*) and noble fir (*A. procera*). They are spoken of collectively as balsam fir in the interpretation of their pollen profiles. Spruce pollen was identified as largely that of Sitka spruce (*Picea sitchensis*), although it is possible that Engelmann spruce (*P. engelmanni*) may be represented, especially during the cooler periods of the time recorded by the pollen profiles. In addition to that of conifers, pollen of alder, willow, ericads, grass, composites, cattail, and sedges were noted, while spores of ferns and mosses were present. The non-coniferous pollen, however, was scanty, and probably largely destroyed by the compaction of the sediments and the harsh preparation methods. In the upper stratum at Possession Point, leaves of sphagnum moss were noted, but in general, the vegetative structures were indistinguishable due to crushing by the overlying sediments and the rigorous treatment in preparation of the samples.

INTERPRETATION OF THE POLLEN PROFILES

According to the stratigraphic relationships (Mackin) the composite sedimentary column of the Possession Point and

Everett sections may be considered to represent three chronological units of an interglacial stage. Interpretation of the pollen profiles, however, can best be treated in five segments, the P1-4 and P5-15 of the Possession Point sediments, and the A, B, and C series of the Everett section. The Possession Point pollen-bearing segments are separated by inorganic strata and thus do not constitute a continuous record of forest succession, while the sharp breaks in the pollen profiles from the Possession Point P5-15 to the A series of the Everett section and in turn from the A to the B series of the latter, indicate that a hiatus also is present between these segments. The B and C series of the Everett sequence apparently record continuous forest succession. The relative chronological position of the recorded forest sequence in each of the sedimentary segments, as well as the continuity of the forest succession of the composite sedimentary column, may be interpreted upon the basis of the postglacial forest succession recorded in many peat sections in the Puget Lowland (Hansen, 1938, 1940, 1941, 1943, 1947a, 1947b). These pollen-bearing sediments undoubtedly record a continuous forest succession from the time of retreat of the Vashon (Late Wisconsin) ice to the present, largely in response to climatic trends and soil development.

In the lowest and oldest pollen-bearing segment, P1-4 of the Possession Point sediments, lodgepole is recorded as the predominant species in adjacent areas, with proportions ranging from 78 to 61 per cent with an average of 68 per cent (fig. 2). White pine with 4 to 25 per cent and an average of 14 per cent, and spruce with 8 to 16 per cent and an average of 13 per cent, are the next most abundant, while other species are sparsely and sporadically recorded. The predominance of lodgepole suggests an early interglacial forest sequence under unstable physiographic and edaphic conditions in the wake of retreating ice. White pine and spruce denote a cool and perhaps moist climate. The almost complete recorded absence of Douglas fir and western hemlock further supports the inference of an initial forest stage under a cool climate and on recently deglaciated, sterile, mineral soil. In ten postglacial peat sections in the Puget Lowland, lodgepole is recorded to an average of 74 per cent in the lowest level, while white pine, the next most abundant, attains an average of 17 per cent (fig. 5). Douglas fir and western hemlock average only 1 per cent. It is believed that lodgepole existed close to the ice front, and was able to invade deglaciated terrain almost immediately in the wake of the retreating ice, and thrive until physiographic and edaphic stability had been attained and sufficient time had elapsed for migration of Douglas fir and hemlock. The last two species, having a greater longevity, being of greater stature, and more tolerant of shade, gradually replaced the initial lodgepole pine forests.

The occurrence of inorganic strata between the lower pollen-bearing segment (P1-4) and the upper (P5-15) in the Possession Point section, denotes a time interval of unknown magnitude between the two forest stages represented in the pollen profiles. The pollen profiles themselves also reflect a chronological hiatus, perhaps several hundred years. The period was long enough, however, to result in a decline of lodgepole from 61 per cent in the uppermost level of the P1-4 series to only 16 per cent in the lowest horizon of the P5-15 series (fig. 2). On the other hand, the balsam fir complex and western hemlock made significant gains, increasing from a few per cent to 46 and 12 per cent respectively in the P5 horizon. Other significant changes include expansions of subalpine fir and mountain hemlock, while white pine and spruce show declines. As the 11 horizons of the P5-15 series were taken at six-inch intervals in a compacted peat stratum, apparently deposited under normal hydrarch succession, they probably represent a continuous forest sequence. The decline of lodgepole and white pine and the rise of hemlock and fir suggest amelioration of the environment, as the influence of the retreating glacier become more remote. The increase of montane and subalpine species does not fit into such a picture, however, unless their increase denotes their presence in ice-free areas at higher altitudes from which pollen drifted down to the accumulating organic sediments. The recorded absence of Douglas fir is somewhat anomalous in the light of postglacial forest succession in the Puget Lowland, in which Douglas fir superseded lodgepole fairly early and became predominant in the region long before western hemlock had begun to expand significantly (figs. 4 and 5). Also the presence of lodgepole and hemlock in abundance in the same region suggests a series of local areas of diverse environment, particularly soil conditions, as hemlock requires humus for its best development, while lodgepole will thrive under poor soil conditions. Their contemporary records may reflect the presence of floodplain areas still under disturbance or mature bog surfaces supporting local forests of lodgepole, and other more edaphically mature areas favorable for western hemlock. With respect to trends, lodgepole and white pine generally increase upward in the section, hemlock expands to a maximum of 20 per cent then declines, while white fir declines to a low of 23 per cent at the top (fig. 2). In general, the abundance of western hemlock and white fir suggests the waning influence of glaciation, white pine and lodgepole pine forests suggest local areas of disturbance and mineral soil, while mountain hemlock and subalpine fir denote favorable local areas possibly at greater altitudes, but in sufficient proximity to be significantly recorded by their pollen.

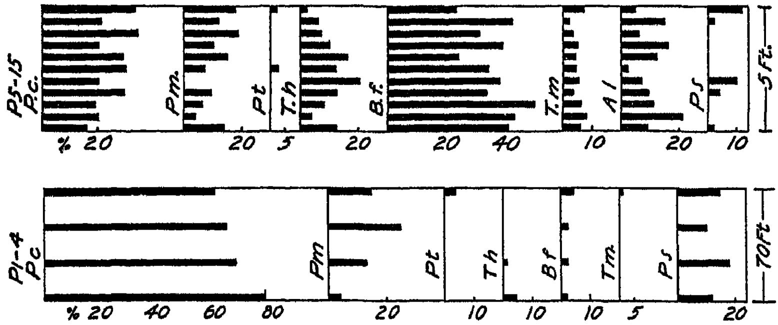

Fig 2 Pollen diagrams of Possession Point section The horizons of the P1-4 series are spaced at greater intervals to indicate greater thickness and longer time interval

According to the stratigraphic relationship (Mackin) the P5-15 segment of Possession Point series, may be in part contemporaneous with or predate the A series of the Everett section. In either case, the pollen profiles of the former may portray a local forest sequence, as suggested by lodgepole and western hemlock co-abundance. The A series may well represent the later stage of a regional Douglas fir-hemlock forest which developed under increasingly stable conditions. This is evidenced by a maximum of 27 per cent for Douglas fir at the lowest horizon and its decline to nothing in the upper level, and high predominance of hemlock with an average of over 52 per cent (fig. 3). If the high proportion of lodgepole pine in the Pl-4 unit represents a regional predominance during the early stages of the recorded interglacial in the Puget Lowland, then sufficient time must have elapsed between the deposition of the Pl-4 levels and the A series to permit the development of a lodgepole-Douglas fir forest and its replacement by Douglas fir-hemlock, the last phase of which is recorded in the A series. The maximum Douglas fir expansion must have occurred during the period represented by the lower, inaccessible portion of the Everett section. This theory is supported by the fact that more than 50 per cent of the pollen identified in the lower part of the A series was that of Douglas fir, but the pollen count was insufficient to warrant a reliable index to adjacent forest composition. That Douglas fir may have attained high proportions during the chronological gap in the record, is further supported by its average maximum of 62 per cent in ten postglacial sections in the Puget Lowland (fig. 5). This suggests that Douglas fir had already passed its peak of interglacial expansion and had been largely replaced by hemlock at the time represented by the A series.

The present successional relationships of Douglas fir and western hemlock in the Puget Lowland reveal that Douglas fir is a subclimax species that thrives and persists as a result of clearing or burning of the climax forest. If a forest is undisturbed by fire, disease, or cutting for five or six centuries, Douglas fir is almost entirely replaced by western hemlock and other climax dominants (Munger, 1940). In postglacial forest succession hemlock was not able to supersede Douglas fir until perhaps about 4,000 years ago, when the climate became cooler and moister after the climatic maximum between 8,000 and 4,000 years ago (fig. 5) (Hansen, 1947). The interval between deposition of the Pl-4 and the A series then, may represent the time involved in the development of Douglas fir predominance and its almost complete replacement by a forest of almost pure western hemlock. In postglacial time, the expansion of Douglas fir to its maximum average of 62 per cent and its decline to a point of hemlock predominance is thought to have required perhaps 15,000 years. It is evident, though, that expansion of hemlock under normal forest succession was deferred by perhaps 6,000 years because of increasing warmth and dryness to a degree that was unfavorable for it. This warm, dry interval was also more favorable for periodic fire, which undoubtedly helped Douglas fir to persist as the predominant species. The entire forest sequence of the present study indicates that a moister climate prevailed which may have limited the extent and frequency of forest fires. This permitted forest succession to reach a climax of hemlock-balsam fir, with spruce, white pine, and lodgepole occupying favorable local sites, while Douglas fir had entirely disappeared from sites within range of pollen dispersal. The total absence of Douglas fir pollen is peculiar, as it seems unlikely that a species as aggressive and hardy as Douglas fir could not have survived on local sites, unless it was entirely destroyed in the region by disease.

A time interval of significant magnitude but probably not so great as above is also indicated between the A and B series of the Everett section, by a disconformity (fig. 1). An unrecorded interval is also indicated by the pollen profiles, chiefly by an increase of balsam fir from 1 per cent at the top of the A series to 43 per cent in the lowest horizon of the B series (fig. 3). Balsam fir maintains high proportions throughout the series, denoting that this expansion was for an appreciable period of time as well as regional. A time lapse of significant magnitude is further supported by an abrupt decline of western hemlock from 49 per cent at the top of the A series to 16 per cent at the bottom of the B series, although there are fluctuations almost as great within the B series from one level to the next. Although hemlock declines from an average of 52 per cent in the A series to about 30 per cent in the B series, it remains more abundant than lodgepole which shows a significant increase with an average of 13 per cent as compared to less than 6 per cent for the A series (fig. 3). Balsam fir averages over 31 per cent and attains a maximum of 50 per cent. Spruce, subalpine fir, and mountain hemlock all show increases in the B series, with averages of 9, 13, and 7 per cent respectively. Again the total absence of Douglas fir pollen is strange, as it seems unlikely that this species was entirely absent in the region. In general, the pollen profiles of the B series denote a slight deterioration of the climate, with greater areas favorable for lodgepole, perhaps representing mature bog surfaces. The significant expansion of the balsam fir complex suggests the expansion of noble and silver fir in montane areas, while grand fir may have prospered in the absence of Douglas fir on floodplain areas.

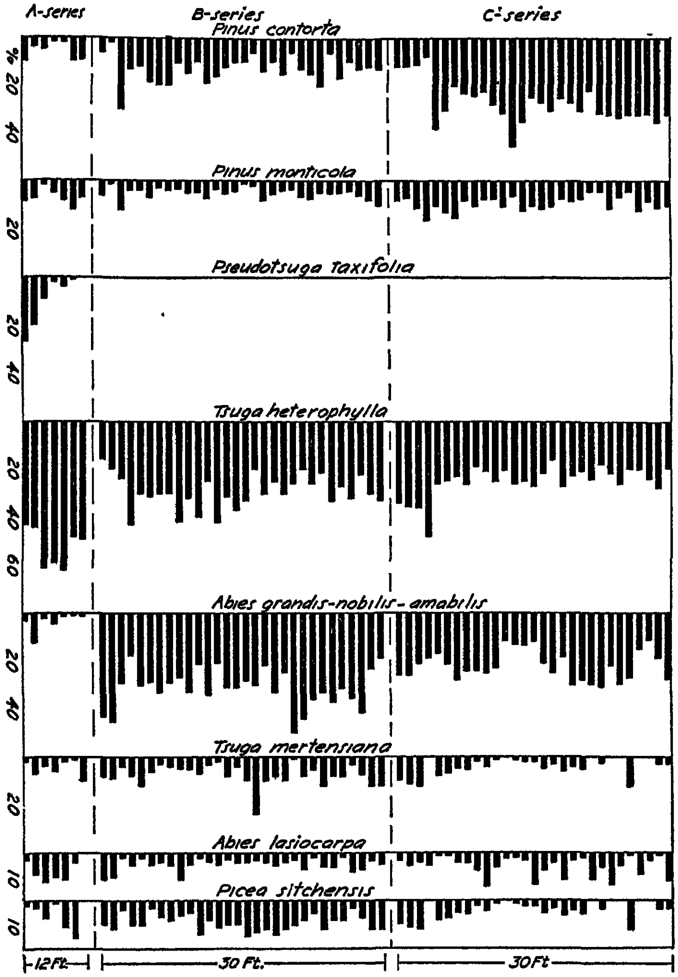

Fig. 3. Pollen diagram of the Everett section. Thickness of pollen-bearing series shown on right of pollen profiles.

The pollen profiles indicate that the B and C series of the Everett section represent a continuous forest sequence. The pollen proportions for all species are almost the same in the upper level of B and lower level of C (fig. 3). During the interim represented by series C, lodgepole gradually increased, hemlock gradually declined, while balsam fir maintained a lower average than in the B series (table 1). White pine also is stronger, with a slightly higher average than below.

The average for lodgepole increases to 27 per cent, its highest since the lower Possession Point unit, white pine increased to 9 per cent, and hemlock declined to 25 per cent and white fir to 19 per cent. The complete absence of Douglas fir pollen is again significant. Spruce, mountain hemlock, and subalpine fir all declined in the C series, recording substantially lower averages (table 1). In general, the pollen profiles of the C series reflect a deterioration of climate and increasing physiographic and edaphic instability, perhaps in response to an approaching glacial influence. The decline of montane and subalpine species may reflect the expansion of mountain glaciers and the continental ice, narrowing the zone available for them. The increase of lodgepole suggests both cooling of the climate and expansion of terrain suitable for its colonization, although a decline of hemlock may also record merely a relative increase in the amount of lodgepole. Also continued development of mature bog surfaces and unstable sand dunes may have favored lodgepole expansion.

As mentioned above, the paucity of pollens in the D series of the Everett section prevents projection of interpreted forest succession above and later than the C series. However, it should be mentioned that in each of the 9 horizons analyzed, over 50 per cent of pollen observed was that of lodgepole, and in one horizon as many as 47 pollen grains were counted, with 65 per cent being lodgepole. Upon this basis, it would seem reasonable to conclude that lodgepole continued to increase upward in the D series, strongly supporting the inference of an accelerating glacial influence.

TABLE I

Ranges and Averages of Pollen Proportions in Per Cent

SPECIES	P1-4 Range	P1-4 Ave.	P5-15 Range	P5-15 Ave.	A series Range	A series Ave.	B series Range	B series Ave.	C series Range	C series Ave
Lodgepole pine ...	61–78	68.0	16–34	24.6	2–10	5.8	2–30	13.1	7–45	27.0
Western hemlock .	0–4	1.2	2–20	8.8	43–62	52.3	16–43	29.4	16–48	25.0
White pine	4–25	14.2	0–19	10.3	2–12	7.0	1–12	5.0	2–18	9.1
Balsam fir . .	2–4	2.5	23–50	36.1	1–12	3.1	18–50	31.4	11–31	19.2
Mountain hemlock	0–1	—	2–8	4.8	1–10	4.5	1–24	7.5	0–12	4.0
Subalpine fir ...	—	—	2–21	10.2	3–12	7.0	2–11	13.3	1–14	5.6
Sitka spruce ...	8–16	13.0	0–12	2.7	1–16	5.7	3–17	9.3	0–12	4.0
Douglas fir .. .	0–4	1.0	0–2	0.5	0–27	9.0	—	—	—	—

CLIMATE AND CHRONOLOGY

The total period of time represented by the composite pollen profiles can be only estimated. The interval as interpreted from the pollen profiles themselves must be estimated upon the basis of the forest succession portrayed and the time required for it, as measured by postglacial forest succession and present-day successional relationships of the species concerned. The thickness and type of organic sediments also provide a clue as to the chronology denoted by the recorded forest sequences. The depositional rate of postglacial pollen-bearing sediments

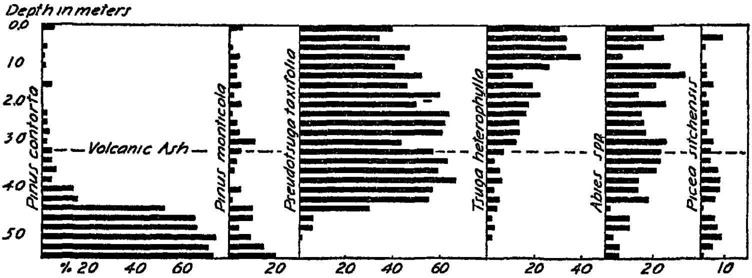

Fig. 4. Pollen diagram of postglacial peat section from bog near Granite Falls, about 15 miles northeast of the Everett section Pollen profiles are typical for postglacial forest sequences in the Puget Lowland.

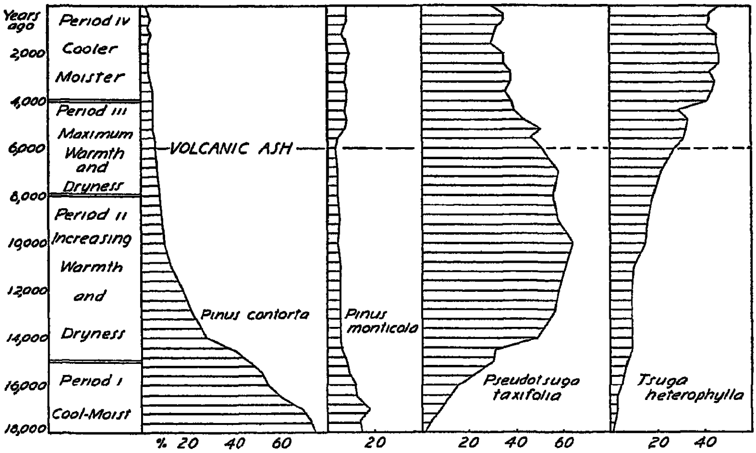

Fig 5 Average pollen profiles of four principal forest trees from ten Puget Lowland postglacial sedimentary columns, most of them within a radius of 50 miles from the Everett section of this study.

may serve as a yardstick, because they are probably a continuous depositional series for a time interval which can be fairly reliably estimated. The texture, structure and thickness of the interbedded inorganic strata also denote relative rates and time required for deposition. The chronological gaps that may exist between the sedimentary units can be estimated by projecting a complete, hypothetical forest succession during an interglacial stage. It must be largely assumed that such an interglacial forest sequence occurred in response to progressive modification and amelioration of the environment, both edaphically and climatically, to a degree of maximum favorability as the glacial influence became more remote and finally disappeared. With the approach of another glacial stage, deterioration of the climate began, followed by periods of physiographic instability at progressively closer intervals as the ice sheet expanded. How long the climatic maximum persisted and the environment remained generally stabilized is conjectural because of the breaks in the record. At present, the climate of the Puget Lowland would seem to be close to a postglacial maximum as far as the climax species present are concerned, which are essentially the same as recorded in the sediments of this study.

The P1-4 unit from the lower Possession Point section denotes a much longer period of time than the four pollen-bearing horizons would indicate. The samples were taken at 10 to 20 foot intervals from 2-4 inch peat layers in a section of inorganic sediments about 70 feet thick (fig. 1). A recorded decline of lodgepole from 78 to 61 per cent is not indicative of a long time lapse in terms of postglacial forest succession. However, the interbedded inorganic members suggest a period of physiographic instability prolonged by slow ice retreat, perhaps from 2,000 to 3,000 years in duration. The climate was probably cool and moist.

The P5-15 pollen-bearing unit of Possession Point is a 5-foot peat bed of almost pure, fibrous, compacted peat, which represents a long period of deposition with at least local physiographic stability. The dense character of the peat suggests compaction to a thickness of only one-third of that of postglacial peats in the Puget Lowland. The average rate of deposition for 30 postglacial, pollen-bearing sedimentary columns in Washington and Idaho has been computed at about 2,500 years per meter. Applying this rate, an interval of 8,000 to 10,000 years may be represented by the P5-15 peat section. The decline of lodgepole from an average of 68 per cent in P1-4 to only 25 per cent in P5-15, and an increase of western hemlock from 1 to 9 per cent and balsam fir from 2.5 to 36 per cent denote some amelioration of the environment. On the basis of postglacial forest succession an interval of 2,000 years is assigned to the gap between the lower and upper Possession Point pollen-bearing units. However, as the stratigraphic relationships suggest that the P5-15 unit predates or is concurrent with part of the A series, and the pollen assemblage may not depict a regional forest sequence, the interim of 8,000 to 10,000 years can hardly be fitted into the total interglacial chronology.

If it is assumed that a regional forest of Douglas fir largely replaced lodgepole as it did in the Puget Lowland during the Postglacial, then the forest successional gap between P1-4 and the Everett A series might better provide a chronological segment of the interglacial. This method of interpretation requires an estimate of the interval involved for the replacement of lodgepole by Douglas fir, development of Douglas fir and western hemlock co-abundance, and decline of Douglas fir and its complete replacement by hemlock. During the Postglacial, it apparently required about 8,000 years for Douglas fir to attain its maximum after deglaciation and 6,000 years more for hemlock to supersede Douglas fir (fig. 5). As mentioned above, however, the warm dry, postglacial maximum probably deferred normal replacement of Douglas fir by hemlock. On these bases the interval from P1 to the top of the A series is estimated from 10,000 to 12,000 years. The pollen profiles of the A series with an average of 52 per cent for hemlock, would seem to depict the period of maximum warmth and physiographic stability for the entire interglacial stage.

A chronologic hiatus of perhaps 2,000 years may be present between the pollen profiles of the A and B Everett series, primarily denoted by an expansion of balsam fir from almost nothing to 44 per cent. The organic laminae themselves contain a high inorganic fraction, and it can be presumed that deposition was rapid, although there was little or no change in forest composition during this stage. The pollen profiles of the B series include about 30 feet of sediments, with the organic strata interbedded in clays, silts, and sands. A figure of 3,000 years is set for the interval represented by the B series. The relative trends of lodgepole and hemlock suggest the beginning of physiographic and edaphic instability. Increase in spruce, subalpine fir, and mountain hemlock suggests cooler and moister climate, which may reflect an approaching glacial influence.

The C series pollen-bearing strata are interbedded in about 30 feet of sediments, becoming progressively coarser upward. Being of equal thickness as the B series, but of texture and structure signifying more rapid deposition, the time interval is estimated at about 2,000 years. A sharp expansion of lodgepole and decline of hemlock suggest increased physiographic instability over that portrayed by the B series, and is strongly indicative of a more proximal glacial influence, while an increase in white pine connotes a cooler climate. Almost total absence of pollen in sediments above the C series may indicate destruction of forest by approaching ice, both directly and through physiographic influence. However, 100 feet of sediments, some fine, between the C series and the top of the section may represent at least 1,000 years of time.

The total time represented by the composite pollen profiles, the inorganic sediments, and the breaks between the several units may be from 16,000 to 20,000 years. Postglacial pollen-bearing sections are estimated at about 18,000 years, which record forest succession from the pioneer lodgepole pine to the present Douglas fir-hemlock forests that existed when white man first appeared. How long the present period of apparent maximum favorability for Douglas fir and hemlock will persist is difficult to say. On the basis of overall trends and long-range cycles, slight deterioration of the climate has been in progress for the last few thousand years, as evidenced by pollen profiles of peat sections from eastern North America and Europe (Hansen, 1947a).

There is no evidence in the several pollen-bearing units of this study for a period of warmth and dryness to the degree reflected in Pacific Northwest postglacial profiles. In general, they suggest a cooler and moister climate throughout the time represented than now exists. The hemlock maximum in the Everett A series probably represents a period of optimum favorability for the development of the climax forest, and the climate of this interval may have been somewhat similar to that of today, but with more moisture. The concurrence of lodgepole and western hemlock throughout, because of their different ecological requirements, suggests either a constant disturbance factor in local areas favorable for lodgepole, such as floodplains, sand dunes, or mature bog surfaces, and stabilized upland areas favorable for hemlock and balsam fir. The sustained high proportions of hemlock throughout suggest that the period represented was an interglacial stage of significant magnitude, rather than a temporary retreat of a minor stage. The stratigraphic relationships show that it was earlier than the interglacial immediately preceding the Vashon (Late Wisconsin) stage in the Puget Lowland.

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RADIUM CONTENT OF ULTRAMAFIC IGNEOUS ROCKS:

II. GEOLOGICAL AND CHEMICAL IMPLICATIONS

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ABSTRACT. Results of recently published measurements of the radium content of some ultramafic igneous rocks and of the constituent minerals are used to demonstrate the relationship of decreasing radium (and uranium) content with increasing mafic character of rocks. The probable redistribution of uranium during partial fusion, crystallization, and alteration is considered. It is shown that in ultramafic rocks the radium is probably present in impurities that represent less than one per cent of the rock and are the consolidation products of the residual solution. The causes of contamination are discussed.

INTRODUCTION

THIS series of studies was undertaken to obtain basic information on the radium and uranium distribution in igneous rocks, particularly in those rocks of ultramafic composition for which few reliable data were available. The results of the laboratory investigation have been presented in this JOURNAL by Davis (1947). Various lines of evidence indicate that the ultramafic rocks make up a large part of the interior of the Earth. The only means available to estimate the radioactivity of the interior and its effect on thermal conditions is by anology with ultramafic intrusives and with meteorites. The results of measurement on some meteorites will be reported in Part III of this series.

There are included here, in the form of an appendix, petrographic descriptions of the rocks examined. This has been done because in examining results of previous investigators there has been considerable uncertainty concerning the identity and history of the materials measured.

Throughout this paper the radium concentrations will be expressed in terms of the customary unit, 10^{-12} gram per gram of rock and uranium concentrations in units of 10^{-6} gram per gram of rock. The uranium in equilibrium with a given radium concentration can be found by dividing the latter by 3.4×10^{-7}.

CLASSIFICATION OF ULTRAMAFIC IGNEOUS ROCKS

For convenience and correlation with previous discussion the ultramafic rocks are here divided into three groups based on the ideas advanced by Hess (1938), which postulated that certain of them were the product of a primary peridotite magma.[1] For practical purposes the division of ultramafic intrusives into three groups as given below is satisfactory regardless of the validity of the hypothesis of an ultramafic magma, since the three have distinctive chemical and mineralogical differences as well as differences in field (genetic) relationships.

1. *Ultramafic Rocks of the Basaltic Magma Suite* (*Crystal Accumulates*). Almost all petrologists are in agreement on the origin of this group of ultramafic rocks which are clearly differentiates of a basaltic magma. In these rocks the settling or accumulation of early formed crystals of olivine, pyroxene, and subordinate chromite to form ultramafic differentiates has been demonstrated in many cases. The ultramafics of the Stillwater and Bushveld complexes are probably the most striking examples. Some specimens from these intrusions have been used for the present investigation.

2. *Ultramafic Rocks of the Secondary Peridotite Suite.* Within the Bushveld and Stillwater complexes, pipes or irregular pseudo-intrusions of ultramafic igneous rocks, commonly dunitic in composition, cut across the ultramafic crystal accumulates. Hess (1938) suggested that these bodies represented magmas formed by partial melting of the crystal accumulates. Bowen's and Tuttle's (1949) more recent explanation that they represent a hydrothermal or pneumatolytic reconstitution of the crystal accumulates by action of volatiles, principally H_2O, seems to be a much more satisfactory explanation. Four determinations on specimens from this group have been made.

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[1] Hess believes that the field evidence requires introduction of these ultramafic igneous rocks as fluid bodies The chemistry which he suggested for such magmas has been demonstrated by Bowen and Tuttle (1949) to be not possible or at best very unlikely Several investigations have suggested the introduction of these bodies by solid flow. It cannot be denied that such material will flow under intense deformation and certain cases of it have been demonstrated. This, however, cannot be a general solution of the problem, since many are intruded at shallow depth into weak and into relatively undisturbed sedimentary rocks The strength of serpentine is somewhat greater on the average than limestone, so that if intrusion by solid flow were a valid mechanism, limestone intrusions would also be common, which is certainly not the case. It would appear that a further search for a chemical basis for an ultramafic magma is in order. The whole problem needs a restatement, which is outside the range of this paper, but will be considered at a later time. While it is assumed throughout this paper that ultramafic magmas did exist, it must be borne in mind that this is an assumption with which certain competent investigators would disagree

3. *Ultramafic Intrusives of the Primary Peridotite Suite.* This is the group of ultramafic intrusives which are found along the major axes of island arcs and alpine mountain belts and which are considered by Hess to represent a primary peridotite magma. The bulk of the specimens examined in the present investigation come from this group. The majority of them are more or less serpentinized.

CONCENTRATION OF RADIUM IN IGNEOUS ROCKS

For purpose of comparison the approximate average values for the radium content of various igneous rock categories are given below in table 1.

TABLE 1

Radium Concentration in Igneous Rocks

	Ra x 10^{-12}g./g.	U x 10^{-6}g /g
Felsic Igneous Rocks	1.0	3.0
Intermediate Igneous Rocks	0.5	1.5
Mafic Igneous Rocks	0.2	0.6
Ultramafic Igneous Rocks*	0.01	0.03
Meteorites	0.001	0.003

* Including serpentinites

The rapidly decreasing content of radium with increasingly mafic character of the rocks is clearly demonstrated. This relationship is a function of the geochemistry of uranium, the parent element of radium. During magmatic differentiation uranium is strongly concentrated in the residual liquids, as indicated by the fact that primary igneous occurrences of uranium minerals are generally associated with granitic pegmatites or hydrothermal veins. Uranium is apparently excluded from the crystal structures of such early formed minerals as olivines, pyroxenes, spinels, plagioclases, etc.[2] The distribution indicated in table 1 is not entirely to be attributed to magmatic differentiation but equally to the more or less reciprocal process of partial fusion by which the primary magmas probably are generated. In this process uranium might be expected to be concentrated in the earlier products of partial fusion, whereas the early formed minerals mentioned above would be among the last to enter the liquid. Thus, for example, a magma generated by the fusion of one-third of the bulk of a basaltic igneous rock might contain virtually all of its uranium but very little of its magnesian silicates.

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[2] Uranium at high temperature will probably have a valence of four (Tomkeieff, 1946). The radius of U^{+4} is given as 0 97A Though its size is such that it might replace Ca^{+2} or Na^{+1} in a structure it requires eight-fold co-ordination with oxygen atoms and no such positions are available in the structures of these minerals

CONCENTRATION OF RADIUM IN VARIOUS CATEGORIES OF ULTRAMAFICS

Three typical examples of ultramafics formed by accumulation of early crystals from a basaltic magma were analyzed for radium and the results appear in table 2 below. The radium concentration is close to 0.01 unit and amounts to about 5% of the concentration to be expected in the average basaltic magma. In view of the location of the parent uranium in these rocks, it may be that an occasional atom of uranium was mechanically incorporated in cavities or imperfections in the crystal structures of the early formed olivine, pyroxene, and chromite. But it seems more probable that most, if not all of it, was contained originally in the small amount of interstitial liquid trapped between accumulated crystals. In an as yet unpublished study of the Stillwater Complex, Hess has found that the interstitial liquid in the crystal mush on the floor of the magma chamber amounts to about 28% of the volume of the mush as it accumulates. So long as interconnecting channels exist between this pore-space liquid of the mush and the overlying magma, diffusion of ions from the pore space to the overlying magma and from the magma to the pore space will take place. This permits continued growth of the crystals with decrease of volume of the pore-space liquid until the interconnecting channels become closed. The point at which this occurs is estimated to be as high as 10% pore-space liquid remaining in some cases to as little as 1% in others, with an average in the neighborhood of 3%. From these very rough figures it may be seen that the amount of uranium in the rocks is of the order of magnitude to be expected if it represented the uranium contained in the final small residuum of trapped basaltic liquid.

TABLE 2

Ultramafic Rocks of the Basaltic Magma Suite
(Crystal Accumulates)

Specimen No	Rock Type	Locality	Wgt % Loss at 1000°C	Ra x 10^{-12}g./g
P-402	Bronzitite	Bushveld Complex	0.5	0.010
P-268	Bronzitite	Stillwater Complex	0.5	0.007
P-267	Harzburgite	Stillwater Complex	5.6	0.014
			Average	0.010

The weight loss at 1000°C. represents essentially the percentage of water present in the original rock. If these figures are used to estimate water content they may be slightly low as a result of some increase in weight from oxidation of ferrous iron. The radium content of the rocks of table 2 is very low but is about twice that of those rocks of the Primary Peridotite Suite (table 4) of comparable water content.

The results for four specimens from the so-called Secondary Peridotite Suite given in table 3 average about the same in radium content as those of the group already discussed—the crystal accumulates. The process by which this group was derived from the primary ultramafics does not seem to have affected the radium content very much. In the four specimens analyzed there is a direct correlation between Mg:Fe ratio of the olivine and radium content. It is rather surprising that the more iron-rich the olivine the lower is the radium in the rock, because in general the parent element uranium is concentrated with the lower melting or later crystallizing phase, in this case the more iron-rich olivine. The explanation probably lies in the fact that these rocks are not crystallization products of a magma, but may represent pneumatolytic alteration products of an already crystalline mass (Bowen and Tuttle, 1949). One might suppose that the streaming of water vapor through the crystalline mass may have removed certain materials, such as SiO_2 and MgO, so that the farther the process went the greater the enrichment in the component stable in this environment, namely Fe_2SiO_4. It would appear also that uranium present in the original mass was preferentially carried away by the vapor or volatile phase, to explain the present decrease in radium content with increasing iron.

TABLE 3

Ultramafic Rocks of the Secondary Peridotite Suite

Specimen No.	Rock Type	Locality	Composition of Olivine	Wgt. % Loss at 1000°C.	Ra x 10^{-12} g./g.
P-403	Dunite	Onverwacht Pipe	Fo_{52}	0.5	.0025
P-400	Dunite	Mooihoek Pipe	Fo_{70}	0.5	.006
P-401	Dunite	Driekop Pipe	Fo_{81}	0.5	.0097
P-364	Dunite	Stillwater Complex	Fo_{82}	4.6	.015
P-400, 401, 403		Bushveld Complex		Average	.009

TABLE 4

Radium Content of Rocks of the Primary Peridotite Suite

			% Wgt.	
men No.	Rock Type	Locality	1000°C.	g./g.
P-140	Dunite	Balsam Gap, N. C	0.0	0.0042
P-186-7	Dunite	Balsam Gap, N. C	0.0	0.0031
P-357	Dunite	Dun Mt, New Zealand	0.0	0.002
P-145	Dunite	Twin Sisters, Washington	0.0	0.008
P-391	Dunite	Addie, N. C.	0.8	0.007
P-369	Websterite	Webster, N. C.	0.0	0.003
			Average	0.0046
P-275	Partially serpentinized dunite	Margarita Island, Venezuela	4.5	0.006
P-307	Partially serpentinized dunite	Belvidere Mt., Vt.	6.2	0.004
P-320	Partially serpentinized dunite	Thetford, P. Q.	7.1	0.004
P-318	Partially serpentinized dunite	Belvidere Mt., Vt.	9.7	0.006
			Average	0.005
P-272	Serpentinite	Tagil, Urals, USSR	10.1	0.017
P-324	Serpentinite	Thetford, P. Q.	10.8	0.011
P-270	Serpentinite	Saranovsk, Urals, USSR	11.9	0.020
P-271	Serpentinite	Thetford, P. Q.	12.0	0.013
P-141	Serpentinite	Geiger's Quarry, Octararo Cr., Pa.	13.2	0.016
P-363	Serpentinite	Newfoundland	13.2	0.013
P-273	Serpentinite	Tagils, Urals	14.5	0.020
P-325	Serpentinite	Thetford, P. Q.	14.8	0.019
P-263	Serpentinite	Camaguey, Cuba	16.3	0.017
			Average	0.016

The variation in radium content of the primary peridotite suite, exhibited in table 4, is much smaller than was anticipated in view of the large variation in water content, which ranges from almost zero to more than 16% by weight. As can be seen in table 4 there is a general relationship between water content, as indicated by loss of weight on heating to 1000°C., and radium content. The present number of analyses suggest that it is not a simple relationship. There seems to be a division approximately at 10% weight loss above and below which the relationship is somewhat different. For those in the group with less than 10% loss, the values are all small and show no clear-cut variation. For the group with greater than 10% loss, there appears to be a direct increase of radium with water content if the results are considered by geographic areas (table 5). For the Northern Appalachian Province the results seem to be convincing, but more analyses from other provinces are necessary to establish this point

TABLE 5

Primary Peridotite Suite Arranged by Geographical Division

Specimen No.	Rock Type	Locality	% Wgt. Loss 1000°C.	Ra x 10^{-12} g./g.
		Northern Appalachian Province		
P-307	Partially serpentinized dunite	Belvidere Mt., Vt.	6.2	.004
P-320	Partially serpentinized dunite	Thetford, P Q.	7 1	004
P-318	Partially serpentinized dunite	Belvidere Mt, Vt.	9.7	.006
P-324	Serpentinite	Thetford, P Q	10 8	.011
P-271	Serpentinite	Thetford, P. Q	12.0	.013
P-363	Serpentinite	Newfoundland	13 2	.013
P-141	Serpentinite	Geiger's Quarry, Pa	13 2	.016
P-325	Serpentinite	Thetford, P. Q.	14 8	.019
		Urals Province		
P-269	Chromite Ore	Saranovsk, Urals	—	.016
P-272	Serpentinite	Tagil, Urals	10.1	.017
P-270	Serpentinite	Saranovsk	11 9	.020
P-273	Serpentinite	Tagil	14.5	.020
		Caribbean Province		
P-275	Partially serpentinized dunite	Margarita Is., Venezuela	4.6	006
P-263	Serpentinite	Camaguey, Cuba	16.3	.017

In table 5 is shown an arrangement of results by geographical divisions. The possible causes for a geographic variation indicate several interesting possibilities worthy of further investigation. Some of these are listed below. (1) The variation may be related to a regional variation in the uranium content in the horizon from which primary peridotites are derived. This might be tested by analyzing ultramafics of two or more ages from the same region. (2) It may be related to the per cent of partial fusion of the source rocks in the generation of the parent magma. Such a relation might be established in the resulting ultramafic intrusives by studying their average chemical and mineralogical compositions. (3) There may be a correlation of the radium variation with time, the ages of the intrusions, rather than with geographic distribution as suggested above. The northern Appalachian ultramafics are probably mid-Ordovician, the Urals mid-Carboniferous, and the West Indies Cretaceous.

CONCENTRATION OF RADIUM IN MINERALS SEPARATED FROM ULTRAMAFICS

Minerals were separated from the Addie dunite and Webster websterite to ascertain how radium was distributed among the various solid phases in the rocks. Table 6 shows the results of this investigation.

TABLE 6

Radium in Minerals from Addie Dunite and Webster Websterite

	I Wt Fraction	II Ra x 10^{-12} g /g	I x II Ra contribution to total in rock
WEBSTER Websterite P-369	1 00	007	.007
Olivine	.92	0035	.0032
Chromite	.02	014	.0003
Serpentine	.048	.04	.0019
Tremolite	.012	.12	.0014
Talc	trace	high?	. . .
Kammererite	trace	high?	. .
		Mineral Total .	.0068
WEBSTER Websterite P-369	1 00	003	003
Chrome diopside	80	001	0008
Bronzite	20	002	0002
Minor constituents	trace	high?	.
		Mineral Total .	.0010

In the above rocks the essential minerals, olivine and pyroxene, appear to contain only one-half to one-third of the radium content of the rock. The late-stage minerals present in small amounts, serpentine, tremolite, talc, and kammererite, have relatively high radium contents as might be expected from the previous discussion.

In table 7 the radium values for chromites and their containing rocks are given. Consideration of the chromite from the Balsam Gap dunite, which has seven times as much radium as the rock in which it occurs, might suggest that uranium was able to enter the structure of the chromite crystals. If this were the case, however, we would expect that the radium content of ultramafics would rise with chromite content. Table 7 clearly indicates that this is not the case. As the percentage of chromite in the rock increases, the radium content of the chromite diminishes. In the last example, the chromite ore from Saranovsk, the chromite actually has less radium than the related rock material. The following explanation is offered to account for the radium distribution. Chromite and olivine differ greatly in physical properties. Stresses acting on the rock as a result of crystallization or deformation may tend to produce minute fractures, particularly at the boundaries of two such dissimilar materials. These contacts, therefore, become favored sites for crystallization of the final residuum of the magma, which residuum contains most of the uranium. A somewhat similar situation would exist under the other possibility, namely that the dunites have regenerated olivine during metamorphism. Under either of these hypotheses the amount of uranium present is the amount in the unit volume of magma or rock and its degree of concentration around or in the chromite will be inversely proportional to the amount of chromite present.

Careful examination of the chromite grains in thin section very commonly shows minute kämmererite or chlorite flakes along the outer boundaries of the chromite grains. The coarse chromite from Balsam Gap has fractures in it in which a thin film of white material can be seen with a hand lens. This material is probably kämmererite. It therefore seems possible, if not probable, that the apparent concentration of radium in the chromite of such rocks as the Balsam Gap dunite is in reality related to the presence of a trace of kämmererite which has relatively a very high radium content, and that this late-stage mineral tended to be concentrated at such

TABLE 7

	Ra units in Chromite	Ra units in Rock	Ra Ratio Chromite: Rock	% Chromite in Rock
P-358 chromite, Balsam Gap, N. C.	.028	.004	7	0.5
P-142 chromite, Balsam Gap, N. C. (P-140)	.026			
P-391c chromite, Addie, N.C. (P-391)	014	.007	2	2 0
P-269 chromite in ore,* Saranovsk	.016	.020**	<0.8	85.0

* Contains 15% serpentine
** P-270

physical discontinuities as the boundaries between the silicates and chromite.

Reviewing the distribution of radium in the minerals of the ultramafics, we might suggest as a hypothesis that virtually the entire radium content of the rocks is contained in a small volume of late-stage minerals, particularly the kämmererite, talc, and tremolite, or possibly entirely in one mineral, the kämmererite. Though great care was taken to obtain pure separates of olivine, pyroxene, etc., a tenth of a per cent of an impurity with perhaps 1000 times the radium content of the rock as a whole could easily explain the observed distribution. Such a possibility is suggested by the results obtained on the chromite.

A CONSIDERATION OF THE CAUSES OF RADIUM VARIATION WITHIN THE PRIMARY PERIDOTITE SUITE

The explanation for the very low radium value in the ultramafics with less than 10% water hinges upon the theory of origin of these bodies. If it were assumed that these bodies represent accumulates of early crystals from a basaltic magma (an assumption which the writers are disinclined to make), the uranium would have been expected to have remained in the residual liquid and been concentrated elsewhere just as in the cases of the Bushveld and Stillwater ultramafics. The radium concentration in this group is 0.005 against 0.010 in the Bushveld and Stillwater, so the process would have had to be twice as efficient in carrying away the uranium. If a primary ultramafic magma is assumed, then there still remains two alternative hypotheses for explaining the radium concentration and variation. First, the present low-water content of this group may be a result of a process of differentiation in the ultramafic magma, similar to that in a basaltic magma, by which the early-formed olivine was concentrated at the level at which we see the present exposures, and a volatile-rich residuum carried much of the uranium upward to be ultimately included in the volatile-rich serpentinites at a higher level. This is the hypothesis at present favored. The second alternative is that the present dunites were originally serpentinites, but suffered a high degree of metamorphism during which their volatile content plus certain other constituents including uranium were removed and olivine regenerated. The mosaic textures of most dunites might suggest that they have been metamorphosed, but does not indicate whether they were dunites or serpentinites prior to metamorphism. That olivine will regenerate in serpentinite during metamorphism has been demonstrated by Durrell (1940) and MacDonald (1941).[3] According to the former reference a serpentinite intruded by diorite recrystallized as olivine along the contact, preserving in the magnetite distribution certain of the minor textural features of the serpentinite. No positive choice is possible between these two alternative processes at the present time. Either one or perhaps both may have been effective in bringing about the observed relationships.

Assuming that we are dealing with an ultramafic magma, it is necessary to consider its mode of differentiation. Since it seems likely that the uranium does not enter the early-formed crystals except by accidental mechanical incorporation, it must be concentrated in a residual solution and must finally be trapped in the consolidation product of the residual solution. The early crystals are largely olivine plus a small chromite fraction, followed by the crystallization of subordinate pyroxene. The residual solution becomes enriched in volatiles and possibly enriched slightly in Fe with respect to Mg. Probably after the residual solution reaches the composition of serpentine plus a little magnetite no further differentiation is possible. This is a very limited amount of differentiation as compared to a basaltic magma and probably will not result (as in more common magmas) in a final residuum of small bulk, analogous, for example, to the pegmatites carrying a relatively large percentage of those atoms which cannot enter the crystal structures of common rock-forming silicates. The uranium in this case would be distributed through a large mass of serpentinite which could well represent more than half of the bulk of the original magma.[4]

[3] Better examples have been described by Geoffrey Leech in ultramafic and gabbroid intrusive rocks of the Shulaps, B. C. Unpublished PhD. thesis, Princeton University, 1949.

[4] Small amounts of feldspathic rocks with highly calcic plagioclases are associated with some ultramafic bodies, particularly with those of batholithic proportions Some of these, such as the olivine-anorthite rocks, appear to be differentiates of the ultramafic magma. Others may be intrusions of unrelated younger gabbroic rocks somewhat modified by contact with ultramafic wall rocks. Since such feldspathic rocks are of

RADIUM CONTENT OF THE PERIDOTITE SUBSTRATUM

From the data given in this paper some estimate may be made of the probable radium content of the peridotitic material at the base of the crust. The radium content of all ultramafics examined averages about .010 unit. By direct analogy the peridotites thought to exist at the base of the crust might be assumed to have a radium content of the same order of magnitude. Assuming that the peridotite magma suite is a product of a magma of that composition, a further analysis is possible. This magma would be generated in the lower peridotitic portion of the crust. The relatively anhydrous members of peridotite magma suite average about .005 unit of radium and the hydrous serpentinites contain about .015 unit. The magma from which both components were derived might reasonably contain .006 to .010 unit. It seems likely that in the generation of such a magma by partial fusion, the liquid would contain most of the uranium of any given unit of volume undergoing partial fusion and the residual crystals very little. Thus, the ultramafic magma might have two or three times as much uranium as the material from which it was derived. This would place the radium content of the peridotite substratum at .002 to .005 unit. The possibility that the above reasoning might give figures which are somewhat too high is considered below under contamination.

The radium content of the deeper interior of the Earth will be considered when the results of radium measurements on meteorites are presented in a later paper.

CONTAMINATED SAMPLES AND A CONSIDERATION OF THE CAUSES OF CONTAMINATION

A satisfactory technique was developed for collecting, handling, and preparing the samples so that they were not

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relatively small bulk and comparatively rare, they have not been considered in the above discussion nor have radium determinations been made on them Their origin is at present an unsolved problem

It also seems possible that magmas intermediate between ultramafic and basaltic may exist. These could give rise to the dunite-pyroxenite-gabbro suite in which the volume of the gabbro is of the same order of magnitude as the ultramafic fraction instead of greatly larger as in the basaltic magma differentiation sequence found in the Stillwater and Bushveld complexes. The occurrences in Southeastern Alaska described by Buddington and Chapin (1929) and later by Kennedy and Walton (1946), the Santa Ana Complex of Paraguanao Peninsula, Venezuela, and perhaps the Great Dike of Southern Rhodesia may be of this character contaminated by radium after leaving their original environment. Five samples, however, give radium values higher than average and are suspected of having been contaminated at some time during their histories. In each case evidence of this probable contamination is available either from the thin section examination or the field relations.

The most probable causes of contamination are as follows: (1) contact with or inclusion of wall rocks of much higher uranium content than the magma; (2) introduction of uranium by hydrothermal solutions after solidification of the ultramafic. A possibility of contamination by ground water was also considered particularly since some of the water bound in the serpentine molecule may have been derived from the wall rocks. From what little information is available on the uranium content of ground water and sea water this possibility seems to be very unlikely, since the amount of uranium contained in such water seems to be much too low to have an appreciable effect. Only in the rare case where the magma intruded an oil field would contamination be likely. Unkovskaya (1940) found waters related to oil accumulations to be 5000 times higher in uranium than sea water and petroleum itself to be 50,000 times higher.

Since wall rocks would normally contain 10 to 100 times as much uranium as ultramafics, samples of several contact rocks were tested partly to ascertain whether contamination could result. In table 8 the values for contact rocks are shown. Two of the three samples taken from contacts with wall rocks show appreciable contamination. A fourth sample (P-322) which gave an exceptionally high radium content (.142 unit), was found to contain small xenoliths of chlorite schist, previously overlooked.

In collecting the samples for radium analysis areas of ultramafics cut by pegmatite dikes or areas in which the effects of hydrothermal alteration were evident, were avoided. Three samples were collected from Belvidere Mountain, Vermont. Two of them, P-308 and P-310 above, have probably been contaminated by hydrothermal solutions. The third sample from this locality, P-307, was apparently unaffected by hydrothermal alteration and has a very low radium content. The radium content of P-308 is very slightly high compared to similar rocks of the Appalachian Province and possibly would not have been considered as contaminated except for the unquestioned contamination of a nearby specimen P-310. The very high radium content of the latter plus the presence of introduced pyrite leaves little doubt of the contamination of the sample by post-consolidation hydrothermal solutions.

TABLE 8

Contact Rocks and Contaminated Samples

Specimen No.	Rock Type & Locality	% Wgt. Loss at 1000°C.	Ra x 10^{-12} g./g.	Remarks
P-363	Serpentinite, Newfoundland	13.2	.013	Contact rock, apparently not contaminated.
P-365	Serpentinite, Newfoundland	5.2	.022	Contact rock, probably contaminated.
P-368	Serpentinite, Camaguey, Cuba	14.2	048	Contact rock, contaminated
P-322	Serpentinite, Thetford, P.Q.	12 9	142	Contains chlorite schist xenoliths, contaminated.
P-308	Serpentinite, Belvidere Mt., Vt.	11.7	.023	Cloudy white alteration product and several per cent of chlorite. Probably some hydrothermal contamination.
P-310	Serpentinite, Belvidere Mt., Vt.	9.7	.062	Some alteration product as above and chlorite plus introduced pyrite. Contaminated by hydrothermal solutions.

Throughout the fresh North Carolina dunites, as well as similar rocks from other areas, minute amounts of a colorless amphibole, a chlorite, or kammererite, and in some samples talc, were noted when the thin sections were carefully examined. The amphibole normally forms tiny, euhedral, short, prismatic crystals between larger olivine grains. The chlorite or kämmererite occurs as blades or flakes which form an exceedingly thin film around chromite grains. Some of these blades sharply penetrate adjacent olivine and no doubt have replaced it. These minerals are commonly considered to be of external hydrothermal origin. Near pegmatite dikes fibrous to acicular tremolite replaces considerable volumes of dunite and chlorite and talc are locally abundant. In the latter case, the evidence for external hydrothermal alteration is clear. In the former case, however, the minute amount of similar minerals uniformly distributed through the dunite is considered to be late stage primary minerals by the writers. It must be remembered that these minerals, plus a few per cent of serpentine, contain a large fraction of the radium in the rock and possibly contain all the radium. If they are attributed to an outside source then the original radium content of the unserpentinized ultramafic rocks would be very much lower than the values here given. The relatively small degree of variation in radium content of these rocks seems to the writers to be inconsistent with an external hydrothermal origin. If such were the origin, we would expect a very large amount of fluctuation in radium content of samples of the rocks from one hand specimen to another. The amounts of uranium which might be deposited in the rocks by such a process might well give variations in radium content of several orders of magnitude considering both the original variations in uranium content of the hydrothermal solutions and variation in the amounts of hydrothermal minerals which might de deposited.

APPENDIX

PETROGRAPHIC DESCRIPTIONS OF ROCKS

P-140, Dunite, Balsam Gap, North Carolina. Collected by Hess (HR1). The hand specimen appears to consist entirely of clear, pale green olivine grains 1-2 mm. in diameter. A fraction of a per cent of chromite is present. In the quarry face the olivine is seen to vary greatly in size, some single crystals up to 20 cm. in diameter being noted. In thin section it can be seen that the fine-grained olivine has a granular mosaic texture made up of more or less equidimensional polygons of olivine fitting nicely together with little trace of a matrix. The optic angle of the olivine is within a degree or two of 90° and the sign is positive. The composition has been estimated by Merwin as $Fo_{90}Fa_{10}$. The large olivine grains in thin section show minute (0.01 mm.) flattened chromite crystals scattered regularly throughout. The finer grained, mosaic, olivine does not appear to contain these tiny flattened chromites.

Chromite as sharp euhedral crystals makes up less than 0.5% of the rock. The crystals are about 0.5 mm. or less in diameter. They occur included within the large olivines and scattered among the mosaic olivine. Very commonly they have "shadows" of a chlorite with the following properties — birefringence variable *ca.* 0.01, optically positive, 2V = 45° or larger, portions of the sheaf-like crystals often showing slightly anomalous interference colors in greenish or brownish tints, dispersion of the optic axes not noticeable. Sheaf-like clusters may be found in the large olivines and some of these send long spikes for a millimeter or more out into the olivine. The same chlorite in the mosaic texture areas may show some crystal faces. The total amount of this mineral present is less than 1% The traces of a pink mineral found in the magnetic separation proves to be this same chlorite. X-ray diffraction examination of this mineral by Davis shows it to be identical with the mineral kämmerite. This mineral is generally classified as a chlorite but evidently differs from most chlorites in structure.

A few grains of tremolite with an optic angle of 90° were found in with the mosaic olivine The optic angle suggests an amphibole between tremolite and pargasite. A very thin film of serpentine seems to coat the smaller olivine grains. It also is seen as minute veinlets in the coarse olivine. It has a honey-yellow color and moderate birefringence for serpentine, reaching first order yellow.

P-141, Serpentinite, Geiger's Quarry, Octararo Creek, Maryland-Pennsylvania line. Collected by Hess (HR2). The main mass of the rock is composed of fine flakes of antigorite with pale bluish interference colors. Some areas of "bastite" occur where flakes of the same type are larger and have parallel orientation. A small per cent of olivine relics are present as scattered clusters of small grains having the same optical orientation within a cluster. A network of fine magnetite grains criss-crosses the rock. Magnetite is more abundant around and in the "bastites," and seems to be clearly related to the serpentinization. A few chromite crystals are found in the thin sections. These are about 1 mm. in diameter, and light transmitted through them is reddish-brown. The borders of the grains are ragged and opaque, probably the result of addition of magnetite during serpentinization. Several subhedral crystals of a brass-yellow metallic mineral, probably pyrrhotite, are also present. Minute gash veins of a colorless, almost uniaxial, positive chlorite are evenly distributed through the serpentine. The birefringence is estimated to be about 0.01, and indices of refraction are slightly higher than the serpentine The interference colors are white with a slightly pinkish tinge. The mineral is probably amesite or prochlorite. Probably all of the minerals present are either related to the original crystallization, such as the olivine and chromite, or related to the serpentinization which is considered to be a late stage magmatic alteration in these rocks. Some doubt may be entertained as to the primary origin of the sulfide.

P-145, Dunite, Twin Sisters Mt., Whatcom County, Washington. Collected by T. Thayer, U.S.G.S. The rock is composed of large grains of olivine in crystals several centimeters across. The composition has been estimated by Merwin as $Fo_{92}Fa_8$. Perfectly euhedral chromites are included in the olivine (0.05 — 0.5 mm.). These amount to a small part of one per cent of the rock. Only a trace of serpentine is present. Rod- or plate-like brown inclusions may be seen under high magnification parallel to the cleavage direction of the olivine. They appear to be slightly pleochroic with greater absorption parallel to their long direction and have indices of refraction higher than olivine. Lines of very minute colorless inclusions can also be seen. These are equidimensional and have a lower index of refraction than the olivine. A sheared area within one olivine crystal shows development of a mosaic texture, possibly cataclastic. Most of the olivine has a slightly undulose extinction.

P-183, Diabase (fine); P-185, Basalt (fine); P-186, Dunite (coarse); P-187, Dunite (fine); P-188, Quartzite (coarse); P-191, Sandstone (fine); P-194, Deccan Trap (fine); P-195, Kimberlite (fine); and P-199, Gabbro-diorite (fine). National Bureau of Standards, Radium Standards, Rock Samples. Rock descriptions, chemical analyses, and radium contents are given in the Minutes of the Committee on Standards of Radioactivity, National Research Council, Washington, D. C., 1938 *et seq.*

P-263, Serpentinized peridotite, Camaguey, Cuba. Collected by Hess (CC4). This rock is almost completely serpentinized but a small percentage of small optically positive olivine relics remain. These relics extinguish simultaneously over areas up to 1 cm. in diameter, indicating the size of the original olivine crystals. They have a pronounced cleavage parallel to (010). A few millimeter grains of almost opaque chromite are present. These evidently were originally anhedral against the olivines, and a pseudomorph of a small olivine is partly included in one of them. Magnetite dust is very finely disseminated through the serpentine, giving it a cloudy appearance. There is no magnetite along the outer rims of the chromite grains. Some magnetite is associated with minute isotropic serpentine veins which transect the rock. A peculiar intergrowth was noted of augite and chromite in which the chromite has forms resembling Japanese characters and is surrounded by the pyroxene. The amount of this intergrowth present in the rock is very small—three or four 2 mm. grains per thin section. Most of the serpentine is isotropic or nearly so. Shreds and also minute asbestiform veins of slightly birefringent serpentine are present in subordinate amounts. The interference colors are bluish-gray.

P-267, Serpentinized peridotite, Stillwater Complex, Montana. Collected by J. W. Peoples (463D7). The rock is a serpentinized peridotite. The common pyroxene in it is a bronzite (En_{86}). Olivine appears as small relics. Most of it has been replaced by serpentine. A small amount of diopsidic pyroxene may be noted as inclusions in the bronzite. Several per cent of plagioclase are present (An_{85}). A half per cent of chromite is also found in the rock. A chemical analysis is given in an unpublished thesis by J. W. Peoples, Princeton University.

P-268, Bronzitite, Stillwater Complex, Montana. Collected by J. W. Peoples (465E3h). Analysis and description: Hess, H. H., and Phillips, A. H., Amer. Min. 25, 276, 1940; and Peoples, J. W. *op. cit.*

P-269, Chromite ore, Saranovsk Urals, USSR. Collected by Hess (25e). About 85% of this rock is chromite which occurs as subhedral to euhedral crystals 0.1 to 3.5 mm. in diameter in a matrix which consists largely of almost isotropic serpentine. The chromite departs from being euhedral by rounding of corners or edges of faces. It is chocolate brown to opaque and is transected by fractures and openings along parting planes filled with serpentine. A little bastite serpentine is present. Small grains of a green isotropic mineral with high relief are seen in the serpentine in some places. These are probably uvarovite.

P-270, Serpentinized peridotite, Saranovsk Urals, USSR. Collected by Hess (25r). The rock is completely serpentinized and has rather abundant fine secondary magnetite as veins and disseminations, and along straight lines which apparently represent former olivine cleavages. A few 1 5 mm. euhedral grains of almost opaque chromite are present, most of which do not have magnetite rims. The main areas of serpentine after olivine have a faint yellowish-green color and, under crossed nicols, low birefringence. They appear to consist of aggregates of serpentine cross-fiber veins with negative elongation of the fibers. In some cases sharp euhedral outlines of original olivine crystals can be made out, set in a matrix of very fine-grained, flaky, colorless serpentine of slightly higher birefringence, antigorite. Chromites are both included in the margins of former olivines and between them. A few flakes of a cloudy brown micaceous mineral are present, probably an altered phlogopite Some of these rim chromite and some are in the "matrix" areas of colorless serpentine. They have strong absorption parallel to the cleavage and moderate birefringence. Some are intergrown with a colorless chlorite A little carbonate is present as rims between olivine pseudomorphs and "matrix" serpentine and around chromite grains. One large bastite crystal is present. It seems to be the same sort of faint yellow-green serpentine as the olivine pseudomorphs but differs in texture.

P-271, Serpentinized peridotite, Caribou Lake, Thetford, P. Q Collected by Hess (Q134). The rock is a typical serpentinite consisting almost wholly of serpentine. The only other mineral present is a small amount of opaque chromite The chromites are for the most part euhedral octahedra varying in size from 0.05 to 0.5 mm. No magnetite dust is present. The serpentine has light gray interference colors and shows a great variety of textural forms. The most common is a felted or reticulate pattern of small flakes, characteristic of antigorite. Large areas of a coarse-grained serpentine with undulose extinction are also present.

P-272, Partially serpentinized dunite, Tagil, Ural Mts, USSR. Collected by Hess (10398h). The rock consists of one-third olivine and two-thirds serpentine, pale yellow in thin section. Minute opaque chromite grains (0.1 mm.) are scattered through the rock. Most of these are euhedral but some are rounded. A few larger chromites 1 mm. in diameter are also present. No magnetite was found. The serpentine forms criss-cross veins or flat lenses between olivine relics. It has moderate birefringence, *ca.* 0.013, and a suggestion of cross-fiber structure with the X-direction across the veins. Wisps of brownish material, possibly limonite stained, are found here and there through the serpentine. There are also small areas in the serpentine of a similar material but which has slightly higher indices of refraction and anomalous interference colors, probably a chlorite. A single grain of augite was found in one thin section.

P-273, Serpentinized dunite, Tagil, Ural Mts., USSR. Collected by Hess (10398e2). This rock is petrographically identical with P-272.

P-275, Dunite, Margarita Island, Venezuela. Collected by Hess (VM9a). The rock consists largely of interlocking grains of olivine, 1-3 mm. in diameter, with irregular grain boundaries. The larger crystals show undulose extinction. Serpentine veinlets transect the olivine irregularly in all directions, making up 20-25% of the rock. The serpentine has low gray-blue interference colors and gives a suggestion of a cross-fiber structure. The X-direction lies across the veins. Opaque chromite occurs as euhedral crystals about 1 mm. or less in diameter A small amount of colorless monoclinic pyroxene is present, optically positive, 2V *ca.* 55° and containing two sets of oriented rod-like or plate-like inclusions, probably parallel to (001) and (100). Some of the inclusions are black and others are brown and translucent.

P-307, Serpentinized peridotite, Belvidere Mt, Vermont. Collected by Davis. This rock is 50-60% serpentinized. It has a peculiar texture. The pyroxenes are in some cases sheared out to highly elongate augen (2 mm.) and the grain size of olivines, (0.2 mm.) suggests a pre-serpentinization mosaic texture. The opaque minerals, both chromite and magnetite, are also strung out in lines along shear planes. The serpentine, a rather coarse-grained antigorite, shows a more or less random orientation of flakes and blades 0.1-0.2 mm. in size, thus appearing to be younger than the shearing. The antigorite blades sharply penetrate the olivine relics in such a manner that the relics are most irregular in outline, unlike the rounded olivine relics so commonly seen in serpentinized peridotites.

The antigorite serpentine is optically positive, occurs as irregular flakes or blades, has low birefringence and has pale blue anomalous interference colors. The olivine has an optic angle of 90°, indicating a composition $Fo_{88}Fa_{12}$. The pyroxene in part is enstatite, optically positive, 2V *ca.* 88°, indicating a composition of $En_{90}Fs_{10}$. Pseudomorphs after pyroxene, distinctly showing its cleavage, are also found. They consist of unidentified cloudy material and are densely crowded with opaque rods and plates. These may originally have been augite crystals. Opaque material makes up about 2% of the rock. The large areas probably have cores of chromite with irregular additions to their borders by late magnetite. These larger opaque areas are invariably surrounded by a rim (0.1 mm.) of isotropic serpentine. Narrow discontinuous opaque veinlets and oriented inclusions in the pyroxene pseudomorphs are probably magnetite No sulfides were noted.

P-308, Serpentinite, Belvidere Mt., Vermont. Collected by Davis. This rock is very similar to P-307 except that serpentinization is very nearly complete. Intense shearing is indicated by the stringing out of opaque minerals along parallel lines and by the augen-like character of the pyroxene pseudomorphs. Most of the pyroxenes have been replaced by serpentine but some of the cloudy unidentified alteration product mentioned in P-307 is seen here also. The same material surrounds some of the rare olivine relics. A colorless chlorite is found to the extent of several per cent in the rock. It has an optic angle of approximately 15°, is positive, and has white interference tints with a pink tinge A few small specks of yellow sulfides are present. Areas of white opaque *alteration are noted in the thin section, viewed by reflected* light.

P-310, Serpentinite, Belvidere Mt., Vermont. Collected by Davis. This rock is very similar to P-307 and P-308 except that it shows more alteration. It contains several large grains of yellow sulfide, aureoles of chlorite around the opaque materials and almost a third of the slide is made up of a cloudy, white-reflecting alteration product.

P-318, Serpentine, Belvidere Mt., Vermont. Collected by Davis. This rock is mineralogically and texturally similar to P-307 but has been serpentinized to a higher degree (87%). Sulfides were not found in the specimen nor does it show the turbid white alteration or marked increase over normal in chlorite as found in the contaminated specimens, P-308 and P-310. Magnetite is somewhat more abundant than in the less serpentinized P-307, magnetite plus chromite amounting to 5% by volume. The chromite is translucent brown and is rimmed by opaque magnetite.

P-320, Serpentinized dunite, King Mine, 855 foot level, Thetford, P. Q. Collected by Davis. The rock consists of olivine in 1 mm. grains transected by a network of serpentine veins. Fifty-five per cent by volume of the rock is serpentine. A few enstatite (En_{92}) crystals are present. The enstatite is rimmed by a narrow band of dusty brown alteration product, probably similar to that found in the Belvidere Mt. specimens. In reflected light this material is white. It has a radiate fibrous structure, the fibers being perpendicular to the edge of the pyroxene. Extinction is inclined to the direction of the fibers. It has high relief compared to serpentine and has interference colors which reach a maximum of reddish yellow. In many cases it becomes darkened by magnetite dust towards its outer margin and may grade into solid magnetite. The wide distribution of this type of alteration around pyroxene through the Vermont-Quebec region suggests that it may be a primary late stage magmatic product. The olivine is highly magnesian, approximately $Fo_{95}Fa_5$. The serpentine adjacent to the olivine is a fine flaky antigorite with faint bluish-gray interference colors. The centers of the veinlets are formed of isotropic colorless serpentine. The chromite is limonite-brown in transmitted light and may or may not be rimmed by magnetite. Chromite appears to be molded around enstatite but may be euhedral against olivine. Magnetite often occurs as minute straight veinlets in olivine crystals but is almost absent in the serpentine veins which appear to be younger and cut across the magnetite. No sulfides were found in the specimen. The texture indicates comparatively little shearing. Some olivines have undulose extinction and the parallel orientation of the magnetite veins in the olivine, in some cases at an angle to the cleavage, suggests deposition on shear zones.

P-322, Serpentinite, King Mine, 855 foot level at contact with "granite," Thetford, P. Q. Collected by Davis. Serpentinization is complete in this rock. The main mass consists of an intergrowth of antigorite of several different habits. A considerable portion appears to be bastite. A few opaque euhedral octahedra, probably of chromite (0.2 mm.) are present. Several rounded spots of strongly pleochroic chlorite occur in the rock. These appear to be foreign material, since chlorites of this type are not ordinarily found in serpentinites. They probably are small xenoliths of schist. The chlorite flakes show a strong preferred orientation within any one of the spots. The optical properties of the chlorite are as follows: indices of refraction considerably higher than antigorite, birefringence high for chlorite about 0.02, optically negative, optic angle small perhaps 20°, X yellow, Y pale green, Z green. The sample came from a location near a contact.

P-324, Serpentinite, King Mine, 1000-foot level, Thetford, P. Q. Collected by Davis. The rock, now 80% serpentinized, was originally a fairly coarse-grained enstatite-bearing dunite. The small olivine relics and the pattern of the serpentine outline original olivine grains up to a centimeter in diameter. The optic angle of the olivine is 90° indicating a composition of approximately Fo_{87} The enstatite is positive with a very large optic angle. Much of it has been replaced by bastite. It is rimmed by the same cloudy alteration product as described in specimen P-320 and magnetite dust or solid magnetite may form an outer border around the alteration.

The chromite, about 4% of the rock, has a dark cherry-red color in transmitted light and forms grains up to 0.5 mm. in diameter. It is rimmed by magnetite. The main mass of the serpentine is a fine flaky antigorite with pale gray interference colors. Some veins of chrysotile traverse the rock. These do not have magnetite cores. Little magnetite dust is present except as mentioned above in association with the enstatite and chromite.

P-325, Serpentinite, King Mine, 1000-foot level, Thetford, P Q. Collected by Davis. The rock is completely serpentinized. It consists largely of almost isotropic or isotropic serpentine. A mesh-like network of veinlets may be seen by the slight color differences in plane light Small subparallel veinlets of more highly birefringent serpentine and magnetite traverse the rock. The centers of these veinlets contain magnetite in isotropic serpentine and the borders consist of the birefringent serpentine with gray-white maximum interference colors. A few 0 2 mm crystals of chromite are present In transmitted light these are limonite-brown and have thin rims of later magnetite.

P-357, Dunite, Mt. Dun, New Zealand Collected by A. L. Day for L. H. Adams The rock consists largely of olivine No material remains for petrographic examination.

P-358, Chromite, Balsam Gap, North Carolina. Collected by Hess. This specimen came from a large crystal about 4 cm. in diameter in the coarse-grained dunite. It was considerably sheared and fractured. A trace of white, micaceous material on the fractures is probably chlorite.

P-363, Serpentinite, Bay of Islands, Newfoundland. Collected by Hess (NA21). The rock is completely serpentinized and consists largely of typical mesh-structure serpentine. This serpentine has its X-direction at right angles to veinlets or flattened lenses and has the following pleochroism: X yellow, Y greenish yellow, Z pale green Between the veinlets or lenses forming the mesh structure are some areas of pale green, nearly isotropic serpentine. Bastite crystals are scattered through the specimen. These are optically positive, 2V large, and X yellow, Y greenish yellow, Z green. Irregular areas 0.1 mm. in diameter of opaque material are found sparingly in the serpentine. Most of these appear to be magnetite rather than chromite A chemical analysis of the rock is given in the AM. JOUR. SCI., 35, 324, 1938.

P-364, Dunite, Stillwater Complex, Montana. Collected by Hess (MB20). A description and chemical analysis is given in the AMER. JOUR SCI., 35, 338, 1938.

P-365, Serpentine, Newfoundland. Collected by Hess (NA14). The rock is a partially serpentinized peridotite. It has a pronounced mosaic texture and a banded appearance resulting from thin sheets of granular, lighter-colored augite in the generally darker-colored, olivine-rich peridotite. The darker color of the olivine portion is largely the result of the presence of much black opaque material, probably magnetite, in the yellowish-brown serpentine which fills closely spaced shear planes. The olivine seems to have been much more susceptible to granulation than the pyroxene. Most of the olivine has a grain size of about 0 1 mm. while most of the pyroxene grains are between 0.5 and several millimeters in diameter. A few small grains of opaque chromite are present. The most of the pyroxene is augite, but bronzite is also present in subordinate amount A small amount of secondary colorless amphibole may be seen replacing augite. This may indicate some contamination of the specimen by later hydrothermal solutions.

P-368, Serpentinite, Santa Clara, Cuba. Collected by Hess (C34). This rock is probably contaminated, having been collected two feet from a contact. There is no evidence in the thin section of secondary minerals to cause contamination except for a little limonite staining. The rock is completely serpentinized and has a well-developed mesh-structure with fairly abundant fine magnetite in the veinlets. Polygonal areas of serpentine are commonly isotropic at the cores and are surrounded by fibrous (chrysotile) serpentine with radial extinction. Some yellowish bastite crystals are present. A few anhedral irregular chromite grains are seen (1-2 mm.) which are brown in transmitted light. A chemical analysis of this rock is given in the AMER JOUR. SCI. 35, 324, 1938.

P-369, Websterite, Webster, North Carolina Collected by Hess. (a) Diopside magnetically separated. (b) Bronzite magnetically separated. (c) Whole rock. The rock consists of chrome diopside (80%) and bronzite (20%) plus a very little chromite. It is completely fresh except for a little greenish stain on cracks in the pyroxenes A trace of tremolite may be seen along shear zones. The rock has a mosaic texture made up of polygonal grains ranging from 0 1 to 1 mm. in diameter. In the hand specimen the diopside is bright chrome-green and the bronzite is a little yellowish-brown. Both minerals are colorless in thin section. The bronzite has a composition approximately $En_{88\,5}Fs_{11\,5}$. It is slightly variable in composition. The optic angle is very close to 90°, most grains being negative but a few are positive. It shows a slight tendency to develop exsolution lamellae parallel to (100). When seen, these are exceedingly fine, perhaps 0.0001 mm., and very few are present compared to those in the Stillwater and Bushveld bronzites. The chrome diopside has a composition equivalent to $Ca_{47.6}Mg_{47.6}Fe_{4.7}$, optic angle + 58°, $Z\Lambda c = 39°$, dispersion weak $r > v$, and $\alpha = 1.672$, $\beta = 1.678$, $\gamma = 1.699$. Analyses of both minerals have been made: one published by Hess and Phillips, Amer. Min. 25, 1940; the other will be published shortly by Hess. These were obtained from the same outcrop but not from the same specimens that were analyzed for radium.

P-391, Dunite, Addie, North Carolina. Collected by Davis. This is a fine-grained dunite with a mosaic texture of polygonal grains of olivine 0.2-1.0 mm. in diameter. It contains about 2% of very black opaque chromite as euhedral to slightly rounded crystals (0.5). A little serpentine is present along cracks between grains. The cracks are so fine that it is difficult to estimate the amount of serpentine but it is probably less than 5% of the rock. A few equidimensional grains of a colorless amphibole were noted interspersed with the olivine. This amphibole has a large optic angle and is optically negative. The beta index of refraction is 1.622 and $Z\Lambda c$ about 18°. It appears to be intermediate between tremolite and pargasite. Spike-like minute plates of a chlorite (kàmmererite) project in some cases from the corner of a chromite crystal into the olivine which they sharply transect. This chlorite appears as a pale pink mineral in the magnetic separates (beta = 1.5825).

Since small albitic and granitic dikes are found in the vicinity of this quarry the possibility of contamination must be borne in mind, particularly since a little talc was found in the rock. The olivine has the following indices of refraction: $\alpha = 1.653$, $\beta = 1.667$, $\gamma = 1.687$, with a composition of $Fo_{91}Fa_{9}$ as estimated by Merwin. The serpentine is antigorite, isotropic with $n = 1.555$ in the form of rough scales. The talc is uniaxial negative with nearly parallel extinction, $\alpha = 1.54$, and $\beta + \gamma = 1.58$, and contains drawn-out brown inclusions.

P-400, Dunite, Mooihoek Pipe, Bushvelt, Transvaal. Collected by E. Sampson (7656). The rock consists largely of coarse (5 mm.) grains of olivine which have irregular interlocking grain boundaries and show banded undulose extinction indicating some strain. The olivine has a beta index of refraction of 1.715, indicating a composition $Fo_{70}Fa_{30}$. The dispersion is fairly strong, $r > v$. It contains opaque plates, probably of magnetite, lying on the (100) plane of the crystals and these plates commonly consist of an aggregate of rods which parallel the *b* crystallographic axis of the olivine. Irregular spots of black opaque material are also found along grain boundaries and less commonly scattered through the olivine A little euhedral to subhedral amphibole, most of which is colorless but some pale green, is found usually between the olivine crystals but also within them. It has $Z \wedge c$ approximately 18° and a positive optical angle of about 80°. There is very little serpentine present, only one veinlet of serpentine-like material. This vein is in part dark green and isotropic but contains some narrow bands of low birefringent material showing pleochroism (X' across the vein, reddish brown, Z' parallel to the vein, green). A little talc and some magnetite are found also in the vein.

P-401, Dunite, Driekop Pipe, Bushvelt, Transvaal. Collected by E. Sampson (7659). The rock consists largely of rounded interlocking grains of olivine (1-2 mm.). The beta index is 1.6915, indicating a composition of $Fo_{81}Fa_{19}$. Some monoclinic pyroxene is present which is slightly pleochroic, Z pale pink, X and Y colorless to pale green, which might easily be mistaken for hypersthene. It is positive with an optic angle near 55°. A trace of greenish-brown serpentine occurs in the cracks in the olivine. A little magnetite is also present in the cracks. Subhedral opaque inclusions (0.05 mm.) are seen in the olivine and are interpreted to be chromite.

P-402, Bronzitite, Malips Drift, Bushvelt, Transvaal. Collected by E. Sampson (7666). The rock consists almost entirely of interlocking subhedral grains of bronzite with a per cent or two of interstitial chrome-augite and a smaller quantity of bytownite. The bronzite has a Mg:Fe ratio of 87.7:12.3 (atomic %), gamma index, 1.6800 ± .0004 The chrome-augite has the following composition (from chemical analysis) $Ca_{45}Mg_{50}Fe_5$ with 1% Cr_2O_3. The plagioclase is $Ab_{58}An_{42}$ ($\alpha = 1.551, \gamma = 1.558$). The rock is perfectly fresh and no other minerals were found in thin section.

The bronzite is very slightly pleochroic from Z green to X pink in section. It has minute exsolution lamellae of diopsidic pyroxene parallel to (100) the optic plane. These are estimated to be less than a micron thick. There are about 200 of them per mm. They make up roughly 5% of the pyroxene. They extinguish about 40° either side of the *c* axis of the bronzite in the plane (010). Chemical analysis of the pyroxene and optical properties are available.

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PLEISTOCENE STRATIGRAPHIC SEQUENCE IN NORTHEASTERN KANSAS

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ABSTRACT. The Pleistocene stratigraphy of northeasternmost Kansas is reinterpreted on the basis of several new exposures in the bluffs of the Missouri River Valley Recognition of early Nebraskan David City gravel, and differentiation of Nebraska and Kansas glacial tills is based primarily on profiles of weathering and regional relationships. The thick late Pleistocene loess section was studied by making a chemical traverse of the deposits and by zonal collecting of the fossil mollusk faunas. Loveland, Peoria, and Bignell members of the Sanborn formation are recognized Each of these members displays a prominent profile of weathering in its upper part.

INTRODUCTION

THE existence of early Pleistocene glacial deposits and late Pleistocene loess in northeastern Kansas has been recognized for many years. Largely because of lack of a well-exposed sequence of distinguishable stratigraphic units and paucity of subsurface data, however, the exact nature of the succession has been unknown. Recently information of much geologic importance has become available through work of the Kansas Highway Department in making several deep cuts. Excavation of Pleistocene deposits was made to secure material for subgrade fill for Kansas Highway 7 at the foot of the Missouri Valley bluff between White Cloud and Sparks in Doniphan County. The new exposures were studied by us during the summer and fall of 1948 and provide the basis for recognition of Pleistocene subdivisions known elsewhere in the Kansas-Nebraska region but not previously known in northeastern Kansas.

The supplemental data for this paper have been drawn from three special studies now under way by the State Geological Survey of Kansas. The first is a study of the late Pleistocene loess faunas by Leonard; the second is research on the ceramic properties of the loesses of northern Kansas by Frye, Plummer, and Runnels; and the third, in cooperation with the Ground Water Division of the United States Geological Survey, is a subsurface investigation of the Pleistocene deposits of the glaciated part of Kansas by use of the State Geological Survey's test drill. Stratigraphic classification presented here will be used in each of these studies.

The terminology of Pleistocene stratigraphic units employed in this paper is that officially adopted by the State Geological

Survey of Kansas. This makes distinction between (1) rock units, such as Yarmouth gravel, which have names consisting of proper nouns; and (2) time-stratigraphic and (3) time units, such as Aftonian Stage and Aftonian Age, which have names consisting of adjectives derived from a proper noun.

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REGIONAL SETTING

The exposures described here occur within a relatively small area along the southwest bluff of the Missouri River Valley in the northeastern part of Doniphan County, the northeasternmost county of Kansas. They are situated adjacent to the Missouri state line, and a few miles south of the southern boundary of Nebraska (fig. 1). The locality is more than 60 miles within the mapped southwestern limits of early Pleistocene ice sheets (Schoewe, 1922, 1930), but a considerably greater distance south of the limits of late Pleistocene ice sheets that terminated in northeastern Nebraska and northwestern and central Iowa. Detailed studies of glacial geology in the region north and northeast of Kansas (Kay and Apfel, 1928; Kay and Graham, 1943; Smith and Riecken, 1947; Flint, 1947) preclude the possibility of the existence in Kansas of glacial tills other than those of Nebraskan and Kansan age, but indicate the existence of outwash floods in the Missouri Valley at the time of the retreat of the several Wisconsinan ice sheets. Although late Pleistocene outwash was not observed in the exposures studied, it is judged to have constituted the source for the extensive loess deposits of the Missouri Valley area.

The limits of glacial advance in Kansas have been established (Schoewe, 1930) in the vicinity of the Kansas River Valley westward from Kansas City, and northward to the Nebraska state line in central Washington County, Kansas, but clear differentiation of Nebraska and Kansas glacial tills has been made at very few places within the State. Differentiation of the two tills is made difficult by their lithologic similarity, the lack of adequate subsurface data, and extensive post-Kansan erosion which has produced a discontinuity of existing deposits in much of the glaciated region.

The distribution of loess in northeastern Kansas is controlled by the location and orientation of the major outwash-carrying valley. In eastern Kansas the thickest deposits of loess occur south of the Missouri River Valley where it swings sharply to the east-southeast in Doniphan County (fig. 1). The greatest thickness of loess known in Doniphan County is 195 feet, penetrated in a test hole drilled in the SE¼ NW¼ sec. 22, T. 1 S., R. 19 E. South of Doniphan County, where the Missouri River Valley trends slightly east of south, the loess is generally less than 25 feet thick, and the upland loess mantle

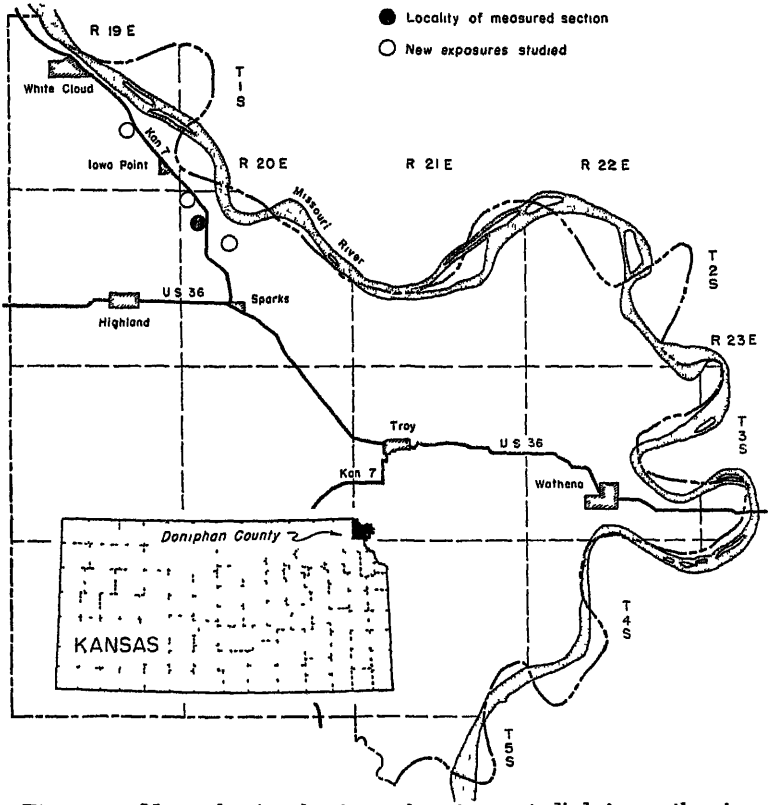


Figure 1 Maps showing location of sections studied in northeastern Doniphan County, and the location of Doniphan County within the State of Kansas.

thins rapidly westward from the valley bluffs. Westward from Doniphan and Brown Counties across the glaciated part of Kansas, the loess is represented only by thin, discontinuous upland remnants. Near the west and southwest glacial limit and 20 to 25 miles east of the Republican River Valley, one again encounters a thick and extensive upland loess mantle which thickens toward its source along Republican Valley.

Inasmuch as profiles of weathering, or fossil soils, are of great importance in classification and correlation of the Pleistocene deposits of northeastern Kansas, a consideration of present climate is pertinent. The nearest stations of the U. S. Weather Bureau are located at St. Joseph, Missouri, a short distance to the southeast in the Missouri River Valley, and at Horton, about 25 miles to the southwest on the uplands. For 60 years of record the average annual rainfall at Horton has been approximately 34 inches, and for a shorter period at St. Joseph, approximately 35 inches. Farther down the Missouri River Valley at Leavenworth, 110 years of record show an average annual rainfall of approximately 35 inches. The mean annual temperature at Horton is 53° F. and at St. Joseph about 1° F. higher. This slight temperature difference is reflected in the average length of the growing season which is 179 days at Horton and 192 days at St. Joseph. In different years the prevailing wind direction at Horton has been from the northwest, southwest, southeast, and south, whereas at St. Joseph it is generally from the south.

STRATIGRAPHY

The most significant among the new exposures studied by us during 1948 is situated in the NE¼ SE¼ sec. 6, T. 2 S., R. 20 E., about 20 miles northwest of St. Joseph, Missouri (pl. 1). At this locality continuously exposed deposits were differentiated and correlated, in ascending order, with the David City formation, Nebraska glacial till, Kansas glacial till, and the Loveland silt, Peoria silt, and Bignell silt, the three last-named being classed as members of the Sanborn formation. Aftonian, Yarmouthian, and Sangamonian interglacial intervals are represented, at least in part, by weathering profiles developed in the upper part of the Nebraska till, Kansas till, and the Loveland silt, respectively. This section presents the most complete record of Pleistocene time known to be exposed at any one locality in Kansas. The succession of strata is described in the following measured section.

Section measured in quarry face and auger holes along Missouri River bluff, NE¼ SE¼ sec. 6, T. 2 S., R. 20 E., Doniphan County, Kansas.

Pleistocene — feet

Sanborn formation

Bignell silt member

Silt, massive, structureless, buff, gray and tan; exposed in vertical face of quarry. Three to four feet at top and bottom partly leached of $CaCO_3$ but containing some etched fossil snail shells. Remainder of interval effervesces in dilute HCI; fossiliferous 38.0

Peoria silt member

Silt, massive, structureless, gray and tan. Well-developed *Brady soil* at top, represented by a few feet gray leached silt, grading downward to a more compact zone, faint reddish buff in color and locally containing abundant, large caliche nodules in lower part; 12 to 17 feet thick. A weakly calcareous zone occurs below the Brady soil and contains fossil snails. Below the fossiliferous zone a few feet of the silt is partly leached and lacks fossil snails. The basal several feet of silt effervesces in dilute HCl and contains fossil snail shells 29.0

Loveland silt member

Silt, massive, reddish-buff. All included in the *Loveland soil* profile, but lower 0.5 foot effervesces in dilute HCl. Upper 7 feet is reddish-buff, partly leached of $CaCO_3$, and displays a joint pattern with $CaCO_3$, and limonite concentrated along some joint planes. In nearby exposures etched fragments of fossil snail shells occur in the lower part of this interval. The sharp boundary at the top of the Loveland and absence of an upper gray layer suggest that part of the Loveland soil may have been removed by erosion prior to the deposition of the overlying Peoria 7.5

Kansas till

Till; matrix of clay and silt containing pebbles and cobbles of limestone, pink quartzite, and igneous rocks. Irregular masses of brown sand are incorporated in the till. Gray and yellow mottled. Lower 7 feet highly calcareous. A post-Kansas pre-Loveland *weathering profile* is indicated by the upper 2 feet of leached till, and the absence of an upper dark layer from the profile caused by pre-Loveland erosion. Large caliche nodules were observed in the lower part of the leached zone and upper part of the calcareous zone; these may represent the lime accumulation zone of the overlying Loveland soil profile 9.0

Nebraska till

Till; matrix of clay and silt with pebbles and cobbles of limestone, igneous rocks, and a few of pink quartzite. Irregular masses of sand and gravel incorporated in till. At top is a well-developed post-Nebraskan pre-Kansan *soil profile* characterized by a black to dark-gray upper zone about 2 feet thick, leached of $CaCO_3$ and lacking limestone pebbles but containing quartzite pebbles and a few igneous rock pebbles. The upper zone grades downward into a medium gray to light brown leached and oxidized zone, which in turn grades downward into gray calcareous till 7.0

David City formation

Gravel; cobbles, boulders, sand, and silt; limestone, igneous rocks, and quartzite. Platy elements roughly shingled. Iron stained, calcareous, locally cemented with $CaCO_3$ in upper part 10.0

Pennsylvanian

Deer Creek limestone

Limestone and shale exposed in quarry face, from level of Missouri River flood plain. Approximately 30.0

Total thickness exposed 130.5

GLACIAL DEPOSITS

The early Pleistocene glacial and glacio-fluvial stratigraphic units are clearly recognizable by lithologic characters observable in the field. The water-laid gravel at the base of the described section is considered to be pro-Nebraskan glacial outwash, which was over-ridden by the advancing Nebraskan ice. Water-deposited gravels, classed as pre-glacial in age, have been observed elsewhere in northeastern Kansas below glacial till (Todd, 1920; Frye, 1941). These pre-glacial gravels consist predominantly of chert pebbles, and lack quartzite and igneous rock types. They reflect a source in the cherty limestones of the Flint Hills, approximately 100 miles west of the Missouri River. In strong contrast, the sub-till gravels considered to be pro-Nebraskan outwash are characterized by rock types similar to those occurring in the calcareous Nebraska till which overlies them. These gravels have not been subjected to prolonged weathering, as is evident from the abundance of unetched limestone pebbles which predominate throughout the deposit. The gravel is here assigned to the David City formation, which was named (Lugn, 1935, pp. 38-40) to include deposits occurring in a similar stratigraphic position in Nebraska. The position of such water-laid gravels resting on Pennsylvanian bedrock approximately 30 feet above the Missouri River flood plain suggests that the present topography has been carved largely since the advent of Nebraskan glaciation.

Above the David City formation is till which may be recognized as comprising two distinct tills by the presence of a well-developed profile of weathering in the middle part of the till section. The lithology of these two tills is quite similar, for they both possess a clay-silt matrix which is highly calcareous below the leached zone in the upper part of each, and both are characterized by the presence of quartzite and igneous rock cobbles and boulders, associated with predominating limestone cobbles and boulders. The only observable lithologic difference between them is a seemingly higher percentage of pink quartzite in the upper of the two tills. The weathering profile in the top of the lower till is several feet thick. This profile is judged to represent a significant interval of weathering, (1) because the relatively impermeable nature of the parent material tends to retard deep weathering, (2) because limestone pebbles are entirely absent from the upper 2 to 2½ feet, and (3) because igneous rocks in this zone are weathered to crumbly incoherent masses. Auger holes bored through the upper till into beds showing this profile prove that the weathered zone is in place and does not represent slump from some higher level.

Correlation of the lower till as Nebraskan in age, and the upper till as Kansan in age is based on field relations to the north and northeast of Doniphan County (Kay and Apfel, 1928; Kay and Graham, 1943; Lugn, 1935) in territory where glacial advances of these two ages—and only these two ages—have been found to occur. The correlation also takes account of the deep profile of weathering between the tills, which excludes the possibility that the two tills studied here represent minor pulsations of one ice sheet. Aftonian time is judged to be represented by the development of the soil profile on the Nebraska till.

A weathered zone also occurs in the uppermost part of the Kansas till, but it is thinner than the Aftonian profile, and lacks the distinctive upper dark layer. These facts suggest that part of the weathering profile on the Kansas till may have been removed prior to the deposition of the overlying loesses. If this be true, the weathering profile on the Kansas till may represent only a part of Yarmouthian time.

In adjacent parts of Kansas and Nebraska water-laid deposits occur between Nebraska and Kansas tills (Holdredge and Fullerton formations of Nebraska classification), unconformably above Kansas till but closely related to it (Meade formation containing Grand Island and Sappa members of Kansas classification), and conformably at the base of the Loveland silt (Crete member, Sanborn formation of Kansas classification).

LOESS DEPOSITS

Criteria for subdivision.—In northern Doniphan County massive silt, or loess, is known to attain a thickness of 195 feet and it constitutes the predominant lithology in the new exposures studied. Seventy-five feet of massive silt was measured above the Kansas glacial till at the locality here described. Superficially, the silt has a similar appearance throughout its entire thickness but subdivision into three correlatable stratigraphic units was made possible by several independent lines of study. The entire section was checked foot by foot with dilute hydrochloric acid in the field. The presence of two major buried profiles of weathering, which were in turn used for field classification, was defined by leached zones determined by acid, by irregular bands of nodular soil caliche, by variation in toughness due to variations in clay content, by slight differences in gross structure, by the range from unetched to etched snail shells, and by the local absence of snail shells. One to two hundred miles west of Missouri

River, in northern Kansas, these fossil soils are more readily recognizable by field inspection owing to the presence of a dark-colored upper layer. The regional differences in morphology of the fossil soils is judged to represent climatic and floral differences at the time of soil formation.

Bulk samples of silt containing fossil snails were collected from the two fossiliferous zones in the Peoria silt, and at regular intervals through the Bignell silt, which is fossiliferous throughout most of its thickness. Fifteen samples of silt were collected for chemical analysis, in order to check the reliability of acid effervescence in the field, and to determine other significant chemical properties of the silt (table 1).

Chemical features.—The chemical analyses, when plotted as vertical traverses, exhibit an unforeseen stratigraphic significance. Figure 2 shows, at the top of each of the two weathering profiles, sharp breaks in the percentage of lime, magnesia, and ignition loss produced by downward leaching during the interval of time these units were not covered by younger deposits. The R_2O_3 content (Al_2O_3, TiO_2, Fe_2O_3) is slightly higher in the upper layer of each fossil soil, suggesting the accumulation of clay. A secondary high within the Peoria falls between the two fossiliferous zones. Some concentration of K_2O may be noted in the upper part of the Loveland profile. The percentages of silica increase in the weathering profiles and decrease in the unweathered zones. Since much of the silica is present in silt-sized grains of quartz, probably it is the most stable constituent of the parent material, and therefore variations in silica content reflect the removal or addition of other more transferrable constituents rather than variations in absolute amounts of silica.

The chemical traverse indicates complexity of the Bignell silt, which is not apparent in the field. The upper part of this member is shown by the analyses to have a distinctly higher calcium carbonate content than the lower part.

Paleontological features.—The species of fossil snails recovered from each of the four fossiliferous zones shown on figure 2 are listed in figure 3. Etched fragments of fossil snail shells occur in the Loveland silt member where it is exposed in a new cut approximately one-half mile northwest of this locality, but identifiable material was not recovered.

TABLE 1

Chemical analyses of loess from locality of measured section in the NE¼ SE¼ sec. 6, T. 2 S., R. 20 E., Doniphan County, Kansas. Stratigraphic position of samples shown by letter on Figure 2. (Analyses by Russels Runnels in the Laboratory of the State Geological Survey of Kansas.)

Sample	SiO_2	Al_2O_3 $+TiO_2$	Fe_2O_3	CaO	MgO	P_2O_5	K_2O	Ign. loss
R	66.58	11.06	3.20	5.60	2.50	0.15	2.84	7.74
P	65.39	10.15	3.25	6.15	2.97	0.15	2.26	8.26
N	68.83	11.09	3.04	4.16	2.41	0.15	2.63	6.10
L	69.11	10.80	3.15	4.22	2.51	0.17	2.52	6.02
K	68.02	10.82	2.76	4.87	3.22	tr.	..	7.00
J	76.99	11.94	2.86	0.98	1.01	0.08	2.21	2.60
I	76.78	11.77	3.82	1.02	0.97	0.10	2.41	2.29
H	77.38	11.82	2.72	0.95	1.26	0.09	2.69	2.41
G	76.79	10.85	2.87	1.10	1.15	0.13	2.58	2.27
F	73.33	12.00	3.02	2.67	1.06	0.08	2.55	3.89
E	71.62	11.61	2.92	3.55	2.08	0.16	2.00	4.68
D	71.21	10.11	2.61	3.70	2.62	0.15	2.35	5.78
C	76.58	12.05	3.03	1.17	1.08	0.17	2.90	2.37
B	76.80	12.05	2.75	0.98	0.96	0.09	2.59	2.48
A	75.63	11.95	2.77	2.17	1.27	0.14	2.64	2.98

The faunal list shown in figure 3 gives only species so far collected from one locality and is not a complete faunal list for the members; studies of loess molluscan faunas are in progress in an area extending across northern Kansas, but this work is incomplete.

The most significant feature of the faunas, as far as present stratigraphic considerations are concerned, is that the fossils in the Peoria silt member at this locality are characteristic of molluscan faunas in the Peoria silt elsewhere in northern Kansas, and especially similar to the molluscan fauna occurring in this loess along the Republican River Valley approximately 125 miles west of this area. The two faunal zones in the Bignell silt coincide remarkably with the change in the chemistry of the silt (fig. 2).

Origin of the loess deposits.—The physiographic and stratigraphic relationships observed in the northeastern Kansas deposits studied, the physical character of the loess, and the contained faunas, the predominant species of which today inhabit the moist lowlands near the water, indicate that the massive silts in Doniphan County were deposited by wind action and were derived from a near-by source on the valley flat of

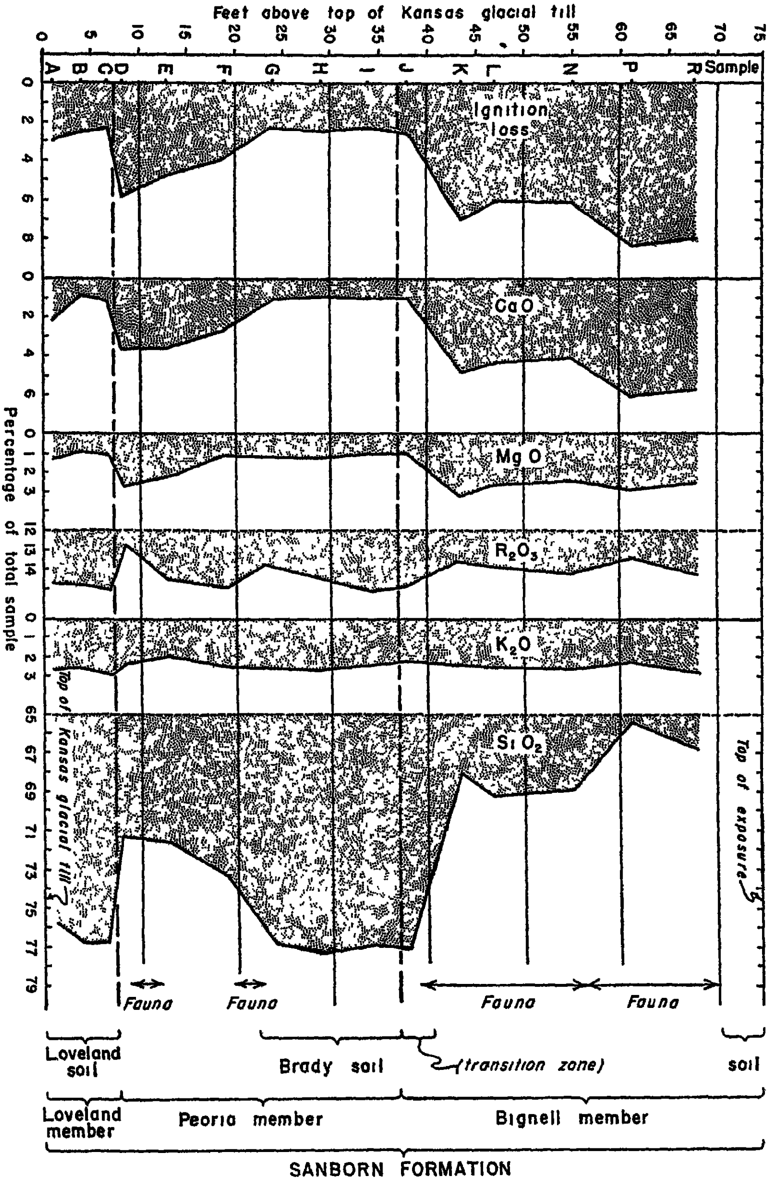


Figure 2 Diagram showing variations in chemical constituents of the three loess members of the Sanborn formation described in the measured section and shown in Plate 1 Analyses by Russell Runnels in the laboratories of the State Geological Survey of Kansas (note table 1). Cemented nodules were not included in the sample for analysis from the caliche zone of the fossil Brady soil.

Species	Peoria (lower zone)	Peoria (upper zone)	Bignell (lower zone)	Bignell (upper zone)
Stenotrema monodon aliciae (Pilsbry and Vanatta)				●
Triodopsis multilineata (Say)				●
Anguispira alternata (Say)			●	●
Hendersonia occulta (Say)			●	●
Gastrocopta armifera (Say)			●	
Helicodiscus parallelus (Say)			●	
Succinea ovalis Say		●		●
Succinea grosvenori Lea		●	●	
Deroceras laeve (Muller)		●		
Discus shimeki Pilsbry		●		
Euconulus sp.		●		
Retinella electrina (Say)		●		
Vertigo milium (Gould)		●		
Discus cronkhitei (Newcomb)	●	●		
Pupilla blandi Morse	●	●		
Succinea avara Say	●	●		
Vertigo tridentata Wolf	●	●		
Hawaiia minuscula (Binney)	●	●	●	
Vallonia gracilicosta Reinhardt	●	●	●	
Total species, each faunal zone	6	13	7	5

Figure 3 Chart showing species of fossil snails obtained from two zones in the Peoria member and two zones in the Bignell member.

the Missouri River. Furthermore, these facts indicate that the most effective wind direction was from the west-northwest. The extreme thickening of loess within the eastward loop described by the Missouri River, the pronounced thinning of the loess mantle within a distance of 20 miles west of the river, and distribution of loess thicknesses in adjacent parts of Missouri, serve as indicators of effective wind direction. That the loess (except locally along steep gully sides) has not moved down slope colluvially is demonstrated by the facts that the fossil soils possess horizontal continuity, are not distorted, and exhibit the physical and chemical characters of weathering *in situ.* It is also proved by the distinctiveness of the four fossiliferous zones, by the existence of the greatest known thickness of loess directly under the highest elements of the local topography, and by the exposure of the top of Pennsylvanian bedrock not far above the flood plain of the Missouri River Valley and the floor of tributary valleys. Water deposition of the loess is precluded by the facts that it occupies the highest elements of local topography, lacks stratification, and possesses a high degree of textural uniformity through a known thickness of 195 feet; that the Bignell displays a gradational contact at the top of the Brady soil; and that aquatic species were not found among the fossil snails in all zones studied, and the fossil shells are not crushed or abraded.

The weathering profiles, although the most important, are not the only secondary features. Small calcium carbonate concretions occur sparsely throughout the loess as a result of the action of percolating ground waters. A more striking effect of ground water may be seen in some exposures where the position of a prolonged stand of the water table is marked by a thin, loosely iron-cemented zone, appearing as a narrow rust-red band. A color contrast in the silt, which may be seen more readily in the field than the important old soils, occurs at this line. Above it, the loess is a uniform tan color, whereas below it, the loess is distinctly grayish, except where iron has been concentrated around former roots, producing conspicuous, more or less vertical, rust-red concretions or streaks, which begin abruptly above and taper off below (pl. 1C). That this change in color and local distribution of iron is related to the water table has been shown by auger borings elsewhere in Kansas made by us through loess into the water table and by the general similarity of the shape of this band to the local topography. In fresh cleavage breaks on vertical bluffs of loess, thin color laminae which may represent annual stands of the water table below the more prominent iron-stained band have been observed. Control of these features by the former position of the water table is suggested by their general conformity in shape to the upper prominent band and their lack of conformity to the position of weathering profiles.

Correlation.—The three stratigraphic units established by the foregoing facts are here correlated with the three widespread late Pleistocene loesses of the Kansas-Nebraska region, classed in Kansas as the Loveland, Peoria, and Bignell silt members of the Sanborn formation (Frye and Fent, 1947; Frye, Swineford, and Leonard, 1948).

Kansas glacial till, which has a weathering profile in the top, immediately underlies the loess, and so it may be presumed that the lowest loess unit in this area cannot be older than

PLATE 1

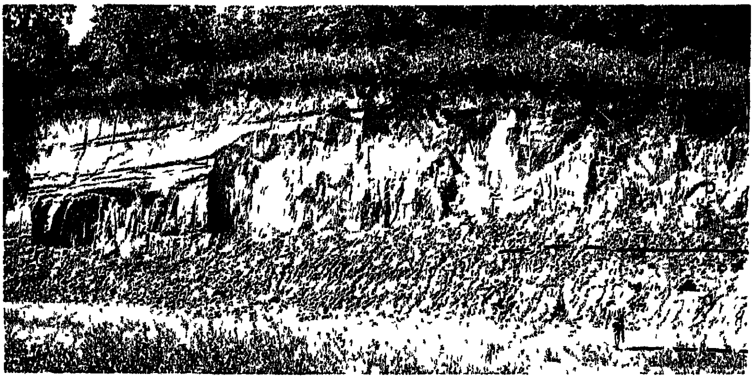
A. Bignell silt member (b) and upper part of Peoria silt member (a) of the Sanborn formation exposed in upper part of highway borrow pit, NE¼ SE¼ sec. 6, T. 2 S., R. 20 E., Doniphan County, Kansas. Man in middle distance is standing on upper fossiliferous zone of the Peoria silt member The stratigraphic section included in this paper was measured at this locality, and the fossil snail faunas listed in figure 3 and the samples for chemical analyses given in table 1 and figure 2 were collected here. (July 1948)

B. Shows method of sampling the Bignell silt member. The Brady soil of the Peoria silt member (a), lower fossiliferous zone of the Bignell silt member (b), and upper fossiliferous zone of the Bignell silt member (c). Same locality as A. (October 1948)

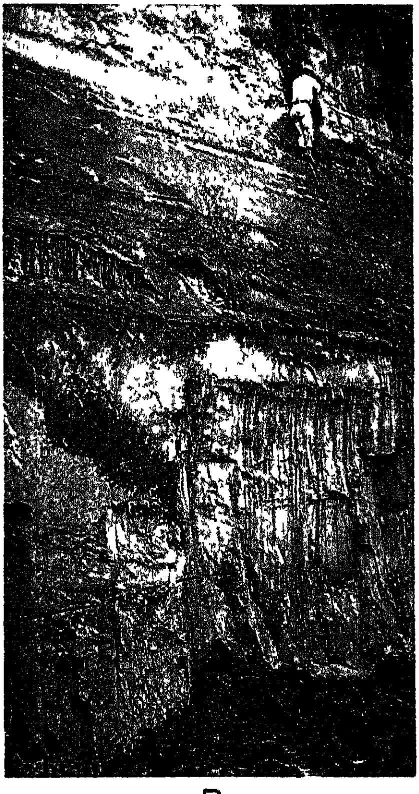
C. Peoria silt member, northeastern Doniphan County, Kansas Former position of the water table is marked by the pronounced iron-cemented band about midway of the hammer handle Above the iron-cemented band the loess is a uniform tan; below, a medium gray with iron cement concentrated in long tubules around root remains. (July 1948)

D. Kansas till at same locality as A. Note pebbles of igneous rock types, and caliche nodule (upper center) judged to represent accumulation from the weathering of the overlying Loveland silt member. (October 1948)

E. Caliche nodules at the base of the Brady soil profile in the upper part of the Peoria silt member, 15 to 17 feet below top of Peoria; 53 to 55 feet below top of exposure. Same locality as A. (October 1948)



A



B

C

D

E

late Yarmouthian or Illinoian. The weathering profile (Loveland soil) and presence of mineralized joints in the Loveland indicate a significant unconformity between the Loveland and the overlying Peoria. The Loveland is the only named loess unit in Kansas and Nebraska known to occupy a stratigraphic position above Kansas till soil and below Peoria loess. It is our judgment that the Loveland silt of this locality was derived from valley train material carried down the Missouri Valley in late Illinoian time.

The second major stratigraphic unit (Peoria silt) within the loess is confined above the Loveland soil and below the top of its contained profile of weathering (Brady soil, (Schultz and Stout, 1945)). It has yielded a molluscan fauna typical of the Peoria silt in central and western Kansas. The Peoria of the Missouri Valley region is judged to be of early Wisconsinan age. As Iowan ice is known to have crossed the present Missouri Valley less than 200 miles north of this area (Flint, 1947), adequate valley train source material probably was available at that time. A minor time break is suggested in the lower part of the Peoria member by the presence of a partial leached zone which lacks a fauna. It is not known, however, whether this should be interpreted as indicating two episodes of deposition during Iowan time, or Iowan deposition followed by post-Iowan deposition.

In north-central Kansas a transition zone commonly occurs at the base of the Peoria silt. This zone has been interpreted as the initial slow accumulation of the eolian deposits of Peorian age at a rate allowing partial leaching contemporaneously with deposition. In the Doniphan County localities studied, such a transition zone at the base of the Peoria is thin or nonexistent, whereas a well-developed transition zone occurs in the basal part of the overlying Bignell silt member.

The uppermost stratigraphic unit, the Bignell silt (Schultz and Stout, 1945; Condra, Reed, and Gordon, 1947; Frye and Fent, 1947), rests on the deep Brady soil profile in the top of the Peoria silt. Therefore, it must be late Wisconsinan and possibly Recent in age. Presumably it was derived from outwash of the relatively near-by Mankato glacial lobe and possibly also from slightly younger or older ice of the Missouri Valley region of South Dakota (Flint, 1947). Although the faunas so far recovered from the Bignell farther west in Kansas are not adequate to permit firm paleontological correlation with the Doniphan County area, the stratigraphic position of the deposit leaves no alternative but to correlate it with this described unit in Nebraska and north-central and western Kansas.

Nowhere in Kansas has a zonation previously been observed in the Bignell silt. Although a weathering profile within it was not recognized in the field, its two-fold nature in Doniphan County is indicated both by the distinctly higher percentage of calcium carbonate in the upper part and the dissimilarity of fossil snail species in its upper and lower zones. The change in faunal elements coincides with the change in calcium carbonate content. These facts, however, in the absence of a well-developed profile of weathering, do not necessarily demand an interval of nondeposition but might be interpreted as the result of an increased rate of deposition in the upper part, a slight change in the nature of source material or local climate, or a change in the cover of vegetation on the surface of sedimentation. The ecological implications of the snail faunas suggest the last of these possibilities. Regardless of which explanation is correct, the vertical change in lithology and fauna imply a change in one or more environmental factors, which, in turn, indicates that the deposition of the Bignell was not sudden but involved an appreciable interval of time.

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STATE GEOLOGICAL SURVEY
THE UNIVERSITY OF KANSAS
LAWRENCE, KANSAS

TWO NEW SPECIES OF ACROCRINUS FROM THE PENNSYLVANIAN OF OKLAHOMA

HARRELL L. STRIMPLE

ABSTRACT. *Acrocrinus pumpkensis*, new species from the Ardmore Basin of southern Oklahoma (lower Pennsylvanian) and *Acrocrinus elegans*, new species from northeastern Oklahoma (upper middle Pennsylvanian) are described.

IN 1940 the author found an interesting specimen of *Acrocrinus*, described here as *A. elegans*, n. sp., in the exposure of shale occurring in the north wall of an abandoned tank dyke some 2½ miles northeast of Copan, Oklahoma. This horizon is some 30′ above the Torpedo sandstone formation, Ochelata group, Missouri series, Pennsylvanian and contains a most interesting and varied fauna. Another specimen of this unique genus was found in the spring of 1948 by a fellow amateur paleontologist, Wm. T. Watkins, in the Pumpkin Creek limestone member, Dornick Hills formation, Pennsylvanian, exposed on the west side of Lake Murray approximately 1½ miles by road from the spillway of the lake, south of Ardmore, Oklahoma. This form proves different from other described species and is presented as *A. pumpkensis*, n. sp.

ACROCRINUS PUMPKENSIS, new species

Plate 1, Figs. 1-3 Text-fig. 1

The dorsal cup is almost as wide at the summit as at mid-height. There are two BB which form a flat disk in the flattened base. The suture between BB extends antero-posteriorly. Stem impression is round and gently concave. There are five RR which are hexagonal shaped plates with the exception of the l. post. R which is pentagonal shaped. Articular facets are concave and do not occupy the entire width of RR. There is a low ridge partitioning the facets into an outer and inner area. The RR are all in lateral contact with the exception of the posterior which are separated by the large anal X.

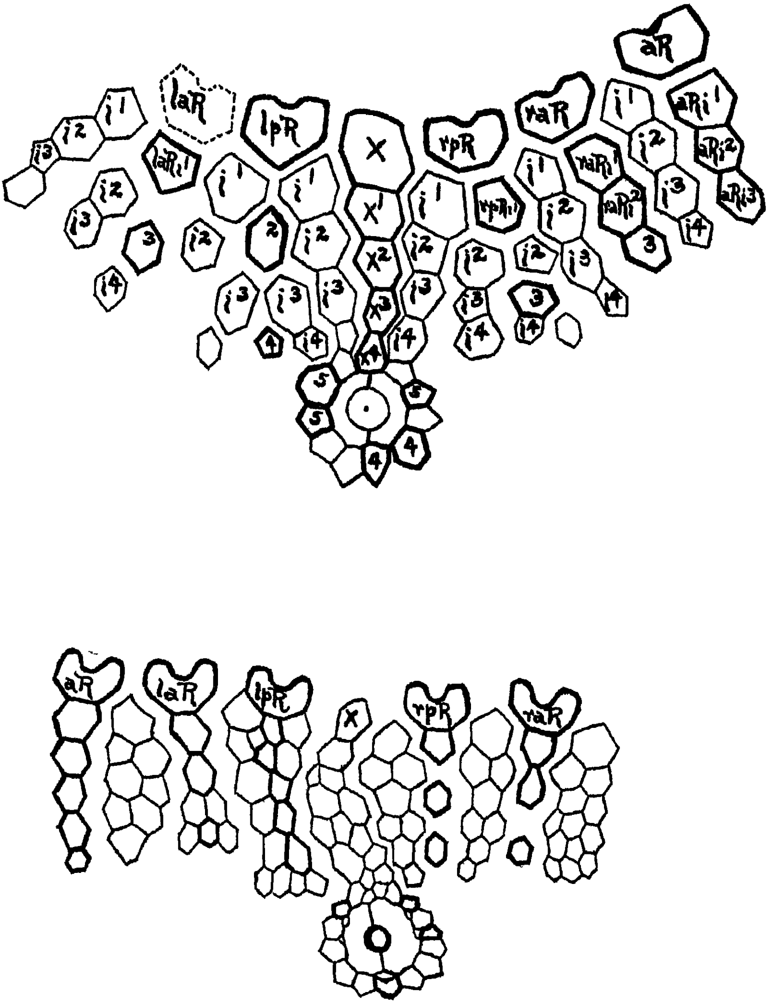
The BB and RR are separated by 4 to 5 circles of plates which are designated as intercalaries (ii) following the terminology of Moore and Plummer, 1937. There are 11 plates in the circlet below the RR and anal X. Thirteen plates may be considered as ii^2; fourteen plates as ii^3; thirteen plates as ii^4; nine plates as ii^5. In the anterior, right anterior and posterior rays, the ii form unbroken series of 4 plates. The left posterior ray is very assymetrical when compared with other series of intercalaries. Most of the ii are hexagonal shaped though many are also of pentagonal outline. Many camerates have symmetrical series of plates in the dorsal cup which are interpreted as being the incorporation of parts of the arm series into the dorsal cup.

Measurements.—	Holotype
Width of dorsal cup (average)	7.5 mm.
Height of dorsal cup	8.4 mm.
Width of ant. R	3.0 mm.
Height of ant. R	1.8 mm.
Diameter of BB circlet	2.3 mm.
Diameter of stem impression	7.0 mm.

Remarks.—Seven species of *Acrocrinus* have been previously described. The genotype species, *A. schumardi* Yandell, and *A. amphora* Wachsmuth and Springer are both Mississippian forms having elongate dorsal cups. *A. praecursor* Springer and *A. intermedius* Springer are also Mississippian species and both have more or less erect, turbinate shaped calices. *A. pirum* Moore and Plummer from the Morrow (lower Pennsylvanian) has a distended basal area and a constricted summit. *A. wortheni* Wachsmuth and *A. brentwoodensis* Moore and Plummer are both Pennsylvanian species having the general calyx outline of *A. pumpkensis*. *A. wortheni* is a smaller form, yet has six circles of intercalaries as compared with the 4 to 5 circles in the Pumpkin Creek species. The complete calyx of *A. brentwoodensis* was not known to Moore and Plummer at the time they presented the species, but the author has a complete dorsal cup from the Hale sandstone which shows the cup outline to be similar to *A. pumpkensis*. In both species anal X is large and is in contact with 3 ii below. In *A. brentwoodensis* the RR are in contact with 4 ii in the three anterior rays and with 3 ii in the two posterior rays. All RR are bordered below by 3 ii in *A. pumpkensis* with the exception of the l. post. R which is supported by only 2 ii.

Occurrence and horizon.—Pumpkin Creek limestone member, Dornick Hills formation, Pennsylvanian; west side of Lake Murray south of Ardmore, Oklahoma.

Holotype.—Collected by Wm. T. Watkins. To be deposited in the U. S. National Museum.



Figures 1 and 2. Diagrammatic sketches of two new species of *Acrocrinus* from Oklahoma. Fig. 1, *Acrocrinus pumpkensis*, n. sp. from the Pumpkin Creek limestone of the Ardmore Basin. Fig. 2, *Acrocrinus elegans*, n. sp. from the shale above the Torpedo sandstone of northeastern Oklahoma.

ACROCRINUS ELEGANS, new species

Plate 1, Figs. 4-6. Text-fig. 2.

The dorsal cup is elongate. Two BB plates are slightly upflared, however the circlet of minute plates surrounding them are almost horizontal in position. There are five RR which are adjoining except at the post. IR where the anal X interrupts the circlet. Articular facets are considerably narrower than the width of RR, and are divided into an outer and inner area by a ridge. There are five to seven circles of intercalaries interposed between the BB and RR. Most ii are hexagonal shaped but several are pentagonal in outline. Each RR is bounded below by 3 ii. It is difficult to determine the exact number of intercalaries to each circlet due to sparodic elements and the disturbed preservation of the post. IR. There are probably twelve ii in the first circle below the RR; possibly 15 to 16 in the second circle; 14 in the third circle; 18 in the fourth circle; 13 to 14 in the fifth circle. Five incipient intercalary plates are present adjacent to the BB, resting in the sutures between larger ii except in the anterior ray where the incipient element is directly under the aRi^6. The plate which may be designated as X^7 also has an immature appearance. Plates of the anterior ray appear to form the only unbroken series in the cup, and those of the left posterior series are the most assymetrical.

The tegmen is in the form of a low tube. A small protruberance is present on the posterior O and is probably a hydropore. Arms are unknown. Under strong magnification all plates have a delicately granular appearance.

Measurements.—	Holotype
Width of dorsal cup (computed)	7.0 mm.
Height of dorsal cup	9.3 mm.
Width of ant. R	2.3 mm.
Height of ant. R	2.3 mm.
Diameter of BB circlet	2.3 mm.
Diameter of stem impression	0.8 mm.

Remarks.—*Acrocrinus elegans* is apparently more closely related to *A. pirum* than other described Pennsylvanian forms. In the later species there is a sharp constriction of the upper portion of the dorsal cup and the lower portion is distended. The species may also be readily separated by the pattern of the intercalaries just below the RR. In *A. pirum* two ii support each RR with the exception of the ant. R which is supported by 3 ii. In *A. elegans* there are three ii adjoining each RR.

The presence of incipient intercalaries adjacent to the BB lends support to the observations of Wachsmuth and Springer, 1897, p. 804; Springer, 1911, p. 41; and Moore and Plummer, 1937, p. 220, wherein it is noted that the intercalaries next below the RR were first to be introduced during ontogenetic development. Irregularity in the shape and placement of ii indicates a primitive reversion toward a cystid-like form. Some authors consider this as the normal evolutionary development among crinoids, however, in the writers' opinion normal evolution is toward simplicity and such a trend as found in *Acrocrinus* is unusual regressive action taken in certain lines.

Occurrence and horizon.—Unnamed shalc 30′ above the Torpedo sandstone formation, Ochelata group, Missouri series, Pennsylvanian; some 2½ miles NE of Copan, Washington County, Oklahoma.

Holotype.—Collected by Harrell L. Strimple. To be deposited in the U. S. National Museum.

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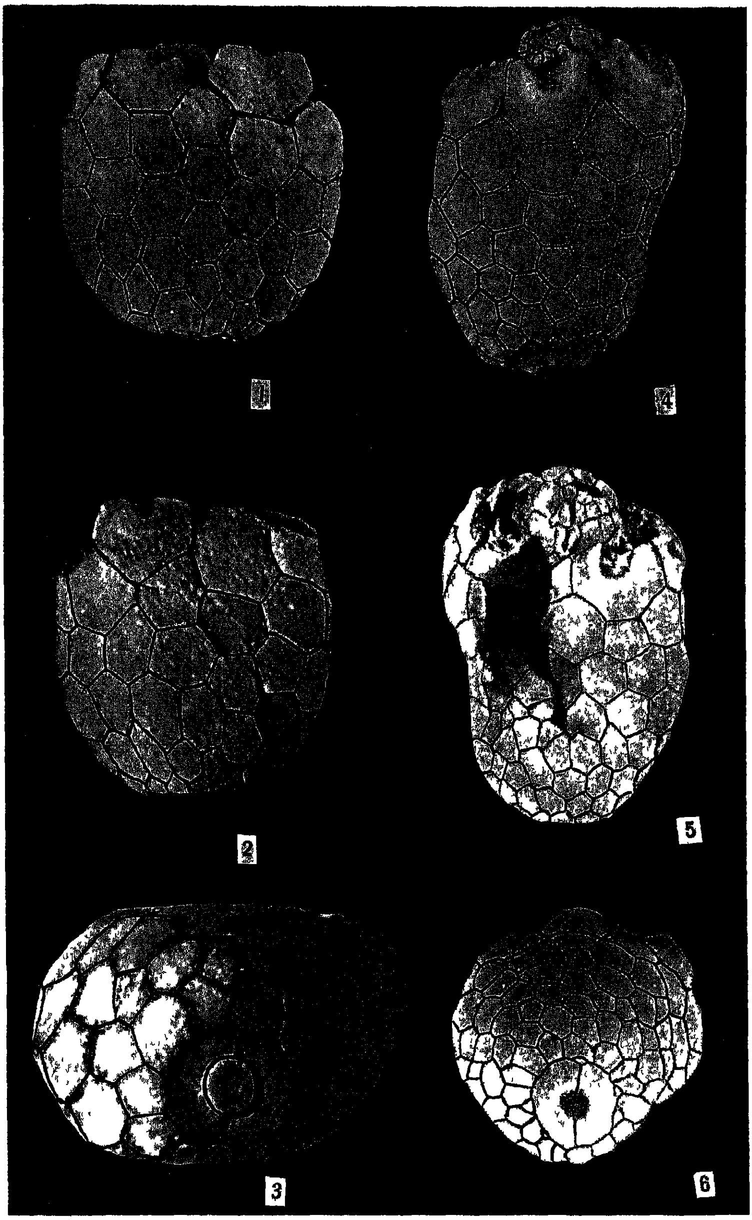

Figs 1-3, *Acrocrinus pumpkensis*, new species 1, Anterior view of holotype, x 4; 2, Posterior view of holotype, x 4; 3, basal view of holotype, x 6

Figs 4-6, *Acrocrinus elegans*, new species 1, Anterior view of holotype, 2, posterior view of holotype, 3, basal view of holotype, anterior radius to the top All views x 4

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INDEX TO VOLUME 247

A

Acrocrinus, two new species from Pennsylvanian of Oklahoma, H. L Strimple, 900.
Adams, Roger, et al, Organic reactions, vol. 4 (Review), 346
Americas, map (Review), 668.
Annual report on the progress of chemistry for 1947, vol 44 (Review), 507.
Anthony, J. W., 791. See Feth, J. H
Arizona volcanics, spheroidal structures, J H. Feth and J. W. Anthony, 791.
Arkose, definitions, S S. Oriel, 824
Atmospheres of the Earth and planets, G. P. Kuiper (Review), 584
Atolls and canyons, Ph. H Kuenen (Review), 830.
Atomic energy, K. K. Darrow (Review), 64.
——— physics, introduction, Otto Oldenberg (Review), 588.
Avian egg, A L Romanoff and A J. Romanoff (Review), 591.

B

Baldwin, Ernest, Dynamic aspects of biochemistry (Review), 506.
Baldwin, R. B., The face of the moon (Review), 585
Barghoorn, E S., and Spackman, William, Jr., A preliminary study of the flora of the Brandon lignite, 33.
Barth, T. F. W., and Rosenqvist, Terkel, Thermodynamic relations of immiscibility and crystallization of molten silicates, 316
Baver, L. D., Soil physics (Review), 70.
Bibliography of the literature on sodium and iodine in relation to plant and animal nutrition (Review), 511.
Biochemistry, dynamic aspects, Ernest Baldwin (Review), 506.
Blackett, P. M. S., Fear, war and the atom bomb (Review), 750.
Blake, D B., and Roy, C J., Unusual stylolites, 779.
Blanckenhorn, Max, obituary of, Rushdi Said, 504.
Bloch, Robert, review by, 589
Bolivia, Andean uplift and erosion surfaces near Uncia, E. H. Walker, 646.
Botanik der Gegenwart und Vorzeit in culturhistorischer Entwickelung, K. F. W. Jessen (Review), 589
Brandon lignite, preliminary study of flora, E. S. Barghoorn and William Spackman, Jr., 33
Brinkmann, Roland, Kayser's Abriss der Geologie (Review), 586.
Briscoe, W L, review by, 65
British Columbia, ultrabasic rocks of Middle River Range, H. W Little, 802.
Brouwer, Dirk, review by, 586.
Brouwer, H. A., Geological explorations in the Island of Celebes; geological summary and petrology (Review), 201.
Brown, S. C, review by, 507.
Brown, T B, Foundations of modern physics (Review), 588.
Buchsbaum, Ralph, Animals without backbones: an introduction to the invertebrates (Review), 670
Burk, R E., and Grummitt, Oliver, Frontiers in chemistry, vol. 5 (Review), 67

C

California, Late Pleistocene and Recent changes of sea level, Santa Barbara County, J. E Upson, 94.
Cambrian, base in southern Appalachians, P. B. King, 513, 622
Canyons, and atolls, Ph. H. Kuenen (Review), 830
Carbon, two aspects of geochemistry, G E. Hutchinson, 27.
Cassidy, H. G., reviews by, 506, 507.
Celebes, geological explorations, summary and petrology, H. A. Brouwer (Review), 201.
Ceramic products; geological interest and analogies, W. O. Williamson, 715
Ceratopsia, growth patterns, R. S. Lull and S W Gray, 492.
Cetaceans, two endocranial casts from Oligocene of New Zealand, B. J. Marples, 462.
Chapman, R. W, and Greenfield, M. A., Spheroidal weathering of igneous rocks, 407.
Chemistry, annual report on progress for 1947, vol 44 (Review), 507.
———, frontiers, vol. 5, R. E. Burk and Oliver Grummitt (Review), 67.
———, general, F E. Ray (Review), 66.

China, metasomatic granitization of batholithic dimensions, part 2. Static, in Sheku area, Peter Misch, 372

Claffy, E. W., The polarographic method for determining trace elements in rocks and minerals, 187.

Clark, A. H, A monograph of the existing crinoids, vol 1 (Review), 205

Cloos, Ernst, review by, 586.

Coastal delineation from aerial photographs, manual, P. G. McCurdy (Review), 286.

Cohen, M. R., and Drabkin, I. E, A source book in Greek science (Review), 351.

Colbert, E. H., A hadrosaurian dinosaur from New Jersey (Review), 68.

Compression tests, oriented specimens of Yule marble, H A. Lepper, Jr, 570.

Corridor of life, W E. Swinton (Review), 206.

Crandell, D. R., review by, 830.

Crinoids, monograph of existing, vol. 1, A H. Clark (Review), 205

Crystal structures, R. W. G. Wyckoff (Review), 350.

Cushman, Joseph, Foraminifera: their classification and economic use (Review), 511

Cybernetics, Norbert Wiener (Review), 342.

D

Daly, R. A, Granite and metasomatism, 753.

Dana's Minerals and how to study them, C. S. Hurlbut, Jr. (Review), 669.

Daniels, Farrington, Outlines of physical chemistry (Review), 143.

Darrow, K. K, Atomic energy; Norman Wait Harris Lectures (Review), 64.

Davis, G. L, and Hess, H. H, Radium content of ultramafic igneous rocks: II Geological and chemical implications, 856.

Dendraster elsmerensis, n. sp, J. W. Durham, 49

Devonian fossils, North American, illustrated catalogue, family Auloporidae, E C Stumm, editor (Review), 205

Dolichoharpes and origin of harpid fringe, H. B Whittington, 276.

Drabkin, I. E, 351. See Cohen, M. R

Dunbar, C O., review by, 205.

Durham, J. W., *Dendraster elsmerensis*, n. sp, 49

E

Earthquakes, J. B. Macelwane (Review), 348.

Electrical measurements in theory and application, A. W Smith (Review), 64.

Enzymes, chemistry and technology, Henry Tauber (Review), 506

Erosion, processes on steep slopes of Oahu, Hawaii, S. E White, 168.

F

Fabric changes in Yule marble after deformation in compression, E. B. Knopf, 433, 537.

Fats, natural, chemical constitution, T. P. Hilditch (Review), 347.

Fault, west edge of Triassic, southern Pennsylvania, G W. Stose, 531.

Fear, war and the atom bomb, P M. S Blackett (Review), 750.

Feth, J. H., and Anthony, J. W, Spheroidal structures in Arizona volcanics, 791

Fix, P. F, Regularity of Old Faithful geyser, Yellowstone National Park, Wyoming, 246.

Floor of the Shenandoah Valley, P. B. King, 73.

Flora of the Brandon lignite, preliminary study, E. S. Barghoorn and William Spackman, Jr, 33.

Foraminifera: classification and economic use, Joseph Cushman (Review), 511.

Frye, A. H., reviews by, 286, 346.

Frye, J C, and Leonard, A. B, Pleistocene stratigraphic sequence in northeastern Kansas, 883.

G

Gallium content of igneous rocks, E. B. Sandell, 40

Garden City formation, stratigraphy and trilobite faunal zones, R. J. Ross, Jr., 472.

Geochemistry of carbon, two aspects, G E Hutchinson, 27.

Geography, human, C L White and G. T Renner (Review), 70.

Geological interest and analogies, ceramic products, W. O. Williamson, 715.

Geologie, Abriss der, Roland Brinkmann (Review), 586.

Geomorphology, textbook, P. G Worcester (Review), 68

Geophysics, Tellus, a quarterly journal (Review), 672

Gesteine und Minerallagerstätten. Vol. 1: Allgemeine Lehre, Paul Niggli (Review), 831.

Gilbert, C. M., and Turner, F. J., Use of the universal stage in sedimentary petrography, 1

Goodman, Clark, The science and engineering of nuclear power, vol. 2 (Review), 588

Granite and metasomatism, R. A. Daly, 753

Granitization, metasomatic, of batholithic dimensions, Peter Misch, 209, 372, 673

Gray, S. W., 492 See Lull, R S.

Greek science, source book, M. R Cohen and I E. Drabkin (Review), 351

Greenfield, M. A , 407. See Chapman, R W

Gregory, J T, review by, 206.

Growth patterns in Ceratopsia, R. S Lull and S W. Gray, 492.

Grummitt, Oliver, 67 See Burk, R. E.

H

Hadrosaurian dinosaur from New Jersey, E. H Colbert (Review), 68.

Hansen, H P., and Mackin, J. H, A pre-Wisconsin forest succession in the Puget Lowland, Washington, 833.

Harpid fringe, *Dolichoharpes* and origin, H B. Whittington, 276

Harrison, G R, Lord, R C, and Loofbourow, J. R., Practical spectroscopy (Review), 344.

Harvard University, Radio Research Laboratory Staff, Very high frequency techniques (Review), 65

Hatai, Kotora, Recent marine shell-bearing Mollusca of the South Sea Islands, part 1 (Review), 671

Hawaii, processes of erosion on steep slopes of Oahu, S E White, 168.

Heat conduction, L. R Ingersoll, et al. (Review), 66.

Hess, H H, 856 See Davis, G. L.

Higgs, D. V, Quantitative areal geology of the United States, 575.

Hilditch, T P., The chemical constitution of natural fats (Review), 347.

Himalayas, metasomatic granitization of batholithic dimensions, part 1. Synkinematic in Nanga Parbat area, Peter Misch, 209.

Historical geology, introduction, R. C Moore (Review), 430

Holmes, C D, National Research Council Committee on the Interrelations of Pleistocene Research (Discussion), 63.

Humphreys, R F., reviews by, 207, 845.

Hurlbut, C S, Jr, Dana's Minerals and how to study them (Review), 669.

Hutchinson, G. E., A note on two aspects of the geochemistry of carbon, 27.

Hybocrinus Crinerensis, new species from Ordovician of Oklahoma, H L Strimple and W. T Watkins, 131

Hypodigms, types and, N D. Newell, 134.

I

Igneous rocks, gallium content, E. B Sandell, 40.

——— ———, spheroidal weathering, R W. Chapman and M. A Greenfield, 407.

——— ———, ultramafic, radium content· II. Geological and chemical implications, G. L Davis and H. H Hess, 856.

Ingersoll, L. R., et al., Heat conduction, with engineering and geological applications (Review), 66.

International Conference on Quaternary Research (Communication), 200

Invertebrates, introduction; animals without backbones, Ralph Buchsbaum (Review), 670.

Isostasy and origin of sial and sima and parental rock magmas, Walter Wahl, 145

J

Jessen, K. F. W., Botanik der Gegenwart und Vorzeit in culturhistorischer Entwickelung (Review), 589

Jones, O. T., and Pugh, W. J., The laccolithic series, 853.

Jones, S. B., review by, 70.

K

Kansas, northeastern, Pleistocene stratigraphic sequence, J. C. Frye and A B. Leonard, 883

Kayser's Abriss der Geologie, Roland Brinkmann (Review), 586

King, P B., The base of the Cambrian in the southern Appalachians, 513, 622.

———, The floor of the Shenandoah Valley, 73

Klockmann's Lehrbuch der Mineralogie, Paul Ramdohr (Review), 668.

Knopf, E. B, Fabric changes in Yule marble after deformation in compression, 433, 537.

Krauskopf, K. B, Fundamentals of physical science (Review), 349.
Krinitzsky, E. L., Origin of pimple mounds, 706
Kuenen, Ph. H, Two problems of marine geology: atolls and canyons (Review), 830.
Kuiper, G P., The atmospheres of the Earth and planets (Review), 584.

L

Laccolithic series, O. T. Jones and W. J Pugh, 353.
Lalicker, C G, Principles of petroleum geology (Review), 751
Langston, Wann, Jr., A new species of *Paleorhinus* from the Triassic of Texas, 324.
Leonard, A B, 883. See Frye, J. C
Lepper, H A., Jr., Compression tests on oriented specimens of Yule marble, 570.
Little, H W., The ultrabasic rocks of Middle River Range, B. C., 802
Liu, S L., review by, 347
Longwell, C. R., review by, 348
Loofbourow, J. R, 344 See Harrison, G. R
Lord, R C., 344 See Harrison, G R
Low, E. M, review by, 511.
Lucke, J. B., review by, 509
Lull, R. S, review by, 68.
———, and Gray, S W., Growth patterns in the Ceratopsia, 492.
Lyons, P. A., review by, 143.

M

Macelwane, J. B., When the Earth quakes (Review), 348.
Mackin, J. H, 833. See Hansen, H. P
Map of the Americas (Review), 668
Margenau, Henry, reviews by, 342, 351, 586, 750, 751.
Marine sediments of southern hemisphere, concentrations of radioelements, W D. Urry, 257
Marples, B J, Two endocranial casts of Cetaceans from the Oligocene of New Zealand, 462
McCurdy, P G, Manual of coastal delineation from aerial photographs (Review), 286.
McKeehan, L. W, reviews by, 64
Metasomatic granitization of batholithic dimensions, Peter Misch, 209, 372, 673
Metasomatism, and granite, R. A Daly, 753.
Microwaves and radar electronics, E. C. Pollard and J. M Sturtevant (Review), 507.
Mineralogical dictionary, Chambers's, E S Smith (Communication), 200.
Mineralogie, Klockmann's Lehrbuch, Paul Ramdohr (Review), 668
Minerals and how to study them, C S. Hurlbut, Jr. (Review), 669
Misch, Peter, Metasomatic granitization of batholithic dimensions, part 1 Synkinematic granitization in Nanga Parbat area Northwest Himalayas, 209; part 2. Static granitization in Sheku area, northwest Yunnan (China), 372; part 3. Relationships of synkinematic and static granitization, 673.
Modern physics, foundation, T. B. Brown (Review), 588.
Mohamed, A F, The distribution of organic matter in sediments from the northern Red Sea, 116
Mollusca of South Sea Islands, recent marine shell-bearing, part 1, Kotora Hatai (Review), 671.
Montgomery, C R, review by, 144
Moon, the face, R. B. Baldwin (Review), 585
Moore, R C., Introduction to historical geology (Review), 430.
Mooreocrinus Bowsheri, new species from Chester series of northeastern Oklahoma, H L Strimple, 128
Morris, P. A, review by, 671.
Morse, P M, Vibration and sound Review), 207.

N

Newell, N D, Types and hypodigms, 134
New Zealand, Oligocene, two endocranial casts of Cetaceans, B J Marples, 462
Niggli, Paul, Gesteine und Minerallagerstatten Vol 1· Allgemeine Lehre von den Gesteinen und Minerallagerstätten(Review), 831
Nuclear power, science and engineering, vol. 2, Clark Goodman (Review), 588.
Nutrition, plant and animal, bibliography of literature on sodium and iodine (Review), 511

O

Obituary of Max Blanckenhorn, Rushdi Said, 504.
Oklahoma, new species of *Hybocrinus Crinerensis* from Ordovician, H. L. Strimple and W. T Watkins, 131.

Oklahoma, northeastern, new species of *Mooreocrinus Bowsheri* from Chester series, H. L. Strimple, 128.
———, two new species of *Acrocrinus* from Pennsylvanian, H. L. Strimple, 900.
Oldenberg, Otto, Introduction to atomic physics (Review), 588.
Old Faithful geyser, regularity, P. F Fix, 246.
Oligocene, New Zealand, two endocranial casts of Cetaceans, B. J. Marples, 462.
Ordovician, Oklahoma, two species of *Hybocrinus Crinerensis*, H. L. Strimple and W. T. Watkins, 131.
Organic chemistry, F. E Ray (Review), 67.
——— ———, essentials, C. D Porter and T. D. Stewart (Review), 286.
——— matter, distribution in sediments from northern Red Sea, A F. Mohamed, 116.
——— reactions, vol. 4, Roger Adams, et al. (Review), 346.
Oriel, S. S., Definitions of arkose, 824.
Origin of sial and sima and parental rock magmas, isostasy, Walter Wahl, 145.

P

Paleorhinus, new species from Triassic of Texas, Wann Langston, Jr, 324
Pennsylvania, southern, fault at west edge of Triassic, G W. Stose, 531
Pennsylvanian, Oklahoma, two new species of *Acrocrinus*, H. L Strimple, 900
Perkins, H. A, College physics (Review), 144.
Petroleum geology, principles, C. G. Lalicker (Review), 751
Pettijohn, F J., Sedimentary rocks (Review), 667.
Physical chemistry, outlines, Farrington Daniels (Review), 143.
——— science, fundamentals, K. B Krauskopf (Review), 349.
Physics, college, H. A. Perkins (Review), 144.
Pierce, Elliott, reviews by, 66, 67.
Pimple mounds, origin, E. L. Krinitzsky, 706.
Pleistocene, Late, and Recent changes of sea level, Santa Barbara County, California, J. E. Upson, 94.
——— research, National Research Council Committee on the Interrelations of, C. D. Holmes (Discussion), 63.
Pleistocene stratigraphic sequence in northeastern Kansas, J. C Frye and A B. Leonard, 883
Polarographic method for determining trace elements in rocks and minerals, E. W Claffy, 187.
Pollard, E. C, and Sturtevant, J. M., Microwaves and radar electronics (Review), 507
Porter, C. D, and Stewart, T. D., The essentials of organic chemistry (Review), 286.
Pratt, W P., review by, 669
Preferred orientation of calcite in Yule marble, F. J. Turner, 593.
Pre-Wisconsin forest succession in Puget Lowland, Washington, H P Hansen and J. H. Mackin, 833.
Publications recently received, 71, 208, 288, 352, 431, 512, 592, 752, 905.
Pugh, W. J., 353. See Jones, O. T.

Q

Quantitative areal geology of United States, D. V. Higgs, 575
Quantum mechanics, L I. Schiff (Review), 751.
Quaternary Research, International Conference (Communication), 200.

R

Radio Research Laboratory Staff, Harvard University, Very high frequency techniques (Review), 65
Radioactivity of ocean sediments VI. Concentrations of radioelements in marine sediments of southern hemisphere, W. D Urry, 257.
Radium content of ultramafic igneous rocks: II. Geological and chemical implications, G L. Davis and H. H Hess, 856.
Rafinesque, C. S., Précis des découvertes somiologiques ou zoologiques et botaniques (Review), 144.
Ramdohr, Paul, Klockmann's Lehrbuch der Mineralogie (Review), 668
Ray, F. E., General chemistry (Review), 66.
———, Organic chemistry (Review), 67.
Recent, and Late Pleistocene changes of sea level, Santa Barbara County, California, J E. Upson, 94.
Red Sea, northern, distribution of organic matter in sediments, A. F. Mohamed, 116
Remington, C L., review by, 144.
Renner, G. T., 70. See White, C. L.

Rice, F O., and Teller, Edward, The structure of matter (Review), 666.
Rodgers, John, reviews by, 201, 349, 667, 668.
Romanoff, A. J, 591. See Romanoff, A. L
Romanoff, A. L, and Romanoff, A. J., The avian egg (Review), 591.
Rosenqvist, Terkel, 316 See Barth, T. F W.
Ross, R. J., Jr., Stratigraphy and trilobite faunal zones of the Garden City formation, northeastern Utah, 472
Roy, C. J, 779 See Blake, D. B.
Rudnick, D, review by, 591.

S

Said, Rushdi, Obituary of Max Blanckenhorn, 504.
Sandell, E. B., The gallium content of igneous rocks, 40
Sanders, J. E., review by, 751
Schiff, L. I, Quantum mechanics (Review), 751
Sclar, C B., review by, 831.
Sea level, Late Pleistocene and Recent changes, Santa Barbara County, California, J. E. Upson, 94
Sedimentary petrography, use of universal stage, C. M Gilbert and F. J Turner, 1
——— rocks, F. J. Pettijohn (Review), 667.
Sharp, R. P., Studies of superglacial debris on valley glaciers, 289
Shenandoah Valley, floor, P. B. King, 73.
Shepard, F. P., Submarine geology (Review), 509.
Silicates, molten, thermodynamic relations, T. F. W. Barth and Terkel Rosenqvist, 316
Smith, A. W., Electrical measurements in theory and application (Review), 64.
Smith, E S., Chambers's Mineralogical dictionary (Communication), 200
Soil physics, L. D. Baver (Review), 70.
Soils, farm, E. L. Worthen (Review), 69
Southern Appalachians, base of Cambrian, P. B. King, 513, 622.
Spackman, William, Jr., 33. See Barghoorn, E. S
Spectra, interpretation, W. M. Venable (Review), 345.
Spectroscopy, practical, G. R. Harrison, R. C. Lord, and J. R. Loofbourow (Review), 344
Sphere of Sacrobosco and its commentators, Lynn Thorndike (Review), 586.
Spheroidal structures in Arizona volcanics, J H. Feth and J. W. Anthony, 791.
——— weathering of igneous rocks, R. W. Chapman and M A. Greenfield, 407
Stewart, T D, 286. See Porter, C. D.
Stose, G. W., The fault at the west edge of the Triassic in southern Pennsylvania, 531
Strimple, H. L., *Mooreocrinus Bowsheri*, new species from the Chester series of northern Oklahoma, 128.
———, Two new species of *Acrocrinus* from the Pennsylvanian of Oklahoma, 900
———, and W T. Watkins, *Hybocrinus Crinerensis*, new species from the Ordovician of Oklahoma, 131.
Structure of matter, F O. Rice and Edward Teller (Review), 666.
Stumm, E. C, editor, Illustrated catalogue of North American Devonian fossils: unit 1-E, part a, family Auloporidae (Review), 205
Sturtevant, J. M., 507. See Pollard, E. C.
Stylolites, unusual, D. B. Blake and C. J. Roy, 779.
Submarine geology, F. P. Shepard (Review), 509
Superglacial debris on valley glaciers, R. P. Sharp, 289.
Swanson, C. L. W., reviews by, 69, 70
Swinton, W. E., The corridor of life (Review), 206.

T

Tauber, Henry, The chemistry and technology of enzymes (Review), 506.
Teller, Edward, 666. See Rice, F. O.
Tellus, a quarterly journal of geophysics (Review), 672.
Texas, new species of *Paleorhinus* from Triassic, Wann Langston, Jr, 324.
Thacher, H C., Jr, review by, 666.
Thermodynamic relations of molten silicates, T. F. W. Barth and Terkel Rosenqvist, 316.

Thomas, H. E., review by, 67
Thorndike, Lynn, The sphere of Sacrobosco and its commentators (Review), 586
Trace elements in rocks and minerals, polarographic methods for determining, E W. Claffy, 187.
Triassic, fault at west edge, southern Pennsylvania, G. W. Stose, 531.
———, Texas, new species of *Paleorhinus*, Wann Langston, Jr, 321.
Trilobite faunal zones, and stratigraphy, Garden City formation, northeastern Utah, R. J. Ross, Jr., 472.
Turner, F J, Preferred orientation of calcite in Yule marble, 593
———, 1. See Gilbert, C. M.
Types and hypodigms, N. D. Newell, 134.

U

Ultrabasic rocks of Middle River Range, B C., H. W. Little, 802
United States, quantitative areal geology, D. V. Higgs, 575.
Universal stage, use in sedimentary petrography, C M. Gilbert and F. J Turner, 1.
Uplift, Andean, and erosion surfaces near Uncia, Bolivia, E H. Walker, 646.
Upson, J E., Late Pleistocene and Recent changes of sea level along the coast of Santa Barbara County, California, 94.
Urry, W. D., Radioactivity of ocean sediments VI Concentrations of the radioelements in marine sediments of the southern hemisphere, 257.
Utah, stratigraphy and trilobite faunal zones of Garden City formation, R J. Ross, Jr., 472.

V

Valley glaciers, superglacial debris, R. P. Sharp, 289.
Venable, W. M., The interpretation of spectra (Review), 345.
Very high frequency techniques, Radio Research Laboratory Staff, Harvard University (Review), 65.
Vibration and sound, P. M. Morse (Review), 207.

W

Waagé, K M., reviews by, 511, 670.
Wahl, Walter, Isostasy and the origin of sial and sima and of parental rock magmas, 145
Walker, E H, Andean uplift and erosion surfaces near Uncia, Bolivia, 646.
Washington, pre-Wisconsin forest succession in Puget Lowland, H. P. Hansen and J H. Mackin, 833.
Watkins, W. T., 131. See Strimple, H L.
Watson, W. W, reviews by, 66, 344, 588.
Whitcomb, Lawrence, review by, 430.
White, C. L, and Renner, G. T., Human geography, an ecological study of society (Review), 70.
White, S. E, Processes of erosion on steep slopes of Oahu, Hawaii, 168.
———, reviews by, 68, 286.
Whittington, H. B, *Dolichoharpes* and the origin of the harpid fringe, 276.
Wiener, Norbert, Cybernetics (Review), 342
Wildt, Rupert, reviews by, 584, 585
Williamson, W O, Ceramic products; their geological interest and analogies, 715
Winchell, Horace, reviews by, 350, 668.
Worcester, P G, A textbook of geomorphology (Review), 68.
Worthen, E. L., Farm soils, their management and fertilization (Review), 69
Wyckoff, R. W. G, Crystal structures (Review), 350
Wyoming, regularity of Old Faithful geyser, P F Fix, 246.

Y

Yule marble, compression tests on oriented specimens, H. A. Lepper, Jr., 570.
———, fabric changes after deformation in compression, E. B. Knopf, 433, 537
———, preferred orientation of calcite, F. J. Turner, 593.

Z

Zoologiques, précis des découvertes somiologiques ou botaniques, C. S. Rafinesque (Review), 144.

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